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NORMAL FAULTS OF THE LAKE CHAMPLAIN REGION[1]

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ABSTRACT

Between the compressional structures of the Green Mountains of Vermont and the dome-like Adirondack mass lie the Lake Champlain lowlands characterized by normal faults, moderate dips, and a scarcity of folds. The normal faults belong mainly to two systems, one having a northeast-southwest strike and one a north-south strike. It appears that both systems were formed simultaneously, and structural evidence seems to show that the normal faulting occurred in Ordovician time, before the Green Mountain overthrusting. These faults cause the pre-Cambrian surface at the base of the Paleozoic series to descend to the east by steps. A study of the stretching involved in the faulting shows that there were tensional forces acting downward toward the east. The distribution and pattern of the faults indicate a primary relation to the Green Mountains and a secondary relation to the Adirondacks, as though due to shear strains between the sinking geosyncline and the up-standing Adirondacks. The normal faults appear to have been formed during the geosynclinal stage, and although conceivably due to unknown magmatic movements, appear to have been formed by the sagging of the geosyncline under the weight of accumulating sediments or to tensional forces which cause the geosynclines.

LOCATION AND STRUCTURAL SETTING

In the northern part of the states of Vermont and New York there are three large elements of structure which, though different in character, appear to be related in origin. The Green Mountains in the

[1] Part of a Ph.D. thesis presented at Harvard University, June, 1931. The writer wishes to thank Dr. H. F. Cleland, who first suggested that this area presented opportunities for investigation; Mr. G. H. Hudson of Plattsburg, New York, for valuable guidance in the field which he has known for many years; Dr. Rudolf Ruedemann for useful advice during the field work; and Dr. K. F. Mather and Dr. Kirk Bryan who gave much valuable advice and aid in the correlation of field data.

east consist largely of Lower Paleozoic sedimentary rocks greatly compressed into north-south–trending folds and overthrust faults.

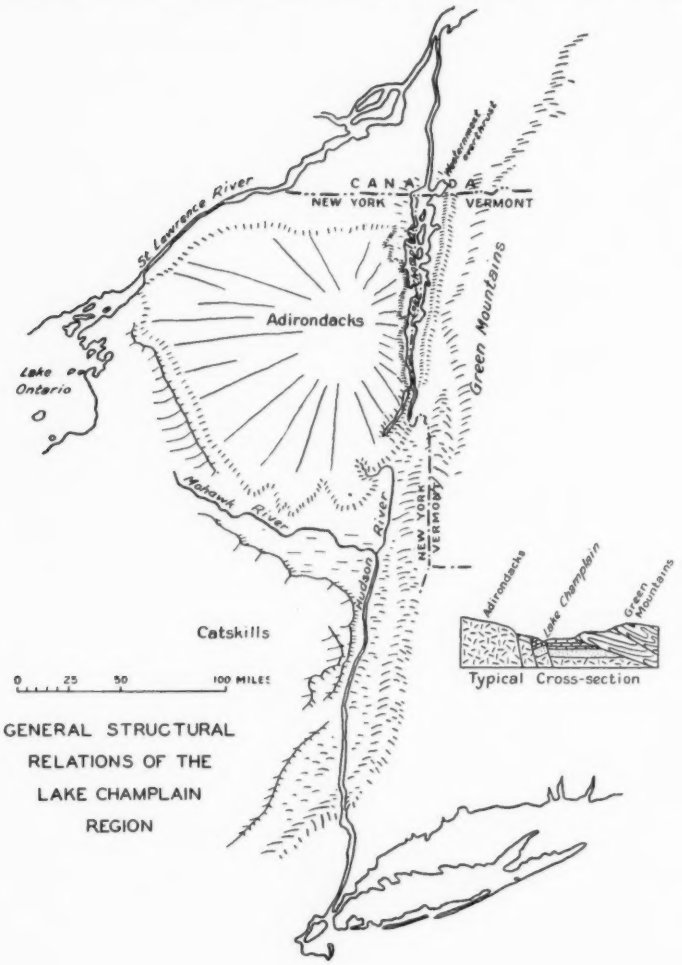


FIG. 1

This area of mountain structure is bordered on the west by the Lake Champlain lowlands which are also composed of Lower Paleozoic

sedimentary rocks. The beds in the lowlands have not been compressed, but are characterized by moderate dips and normal faults. West of the lowland area are the Adirondack Mountains. This mass has a dome-like structure with the Lower Paleozoic beds outcropping around the margins and with the Basement Complex of pre-Cambrian igneous and metamorphic rocks exposed in the interior.

The structure of the Lake Champlain lowlands is the subject of this article. The area covered by field work extends northward from the south boundary of the Ticonderoga quadrangle, near Putnam, New York, to the Canadian line. It is bounded on the east by the westernmost overthrust fault of the Green Mountains and on the west by the pre-Cambrian rocks of the Adirondacks. It comprises the following quadrangles: Mooers, Rouses Point, St. Albans, Dannemora, Plattsburg, Milton, Ausable, Willsboro, Burlington, Port Henry, Middlebury, and Ticonderoga.

STRATIGRAPHY

The rocks of the region are: the pre-Cambrian terranes in the Adirondacks, the Cambrian Potsdam sandstone, the Beekmantown dolomite, the Chazy limestone, a thin series of beds probably equivalent to the Black River and Lowville formations, the Glens Falls limestone of the Trenton group, and Trenton shale of probable Canajoharie age, these last all of Ordovician age.[2] There are also dikes of post-Trenton age[3] and Pleistocene deposits. Space, however, is not available here for a discussion of the stratigraphy; suffice it to say that there are divisions within these formations and that they were

[2] Ezra Brainerd and H. M. Seely, "The Original Chazy Rocks," *Amer. Geologist*, Vol. II (1888), pp. 323–30; "The Calciferous Formation in the Champlain Valley," *Amer. Mus. Natural Hist., Bull. III* (1890), pp. 1–23; "The Chazy of Lake Champlain," *ibid., Bull. VIII* (1896), pp. 305–15. H. P. Cushing, "Geology of the Northern Adirondack Region," *N.Y. State Mus., Bull. 95* (1905). J. F. Kemp and Rudolf Ruedemann, "Geology of the Elizabethtown and Port Henry Quadrangles," *N.Y. State Mus., Bull. 138* (1910). G. H. Perkins, "Geology of Grand Isle," *Vt. State Geologist's Report for 1901–2*, pp. 102–74. P. E. Raymond, "The Chazy Formation and Its Fauna," *Annals of the Carnegie Mus.*, Vol. III (1906), pp. 498–598. Rudolf Ruedemann, "Report on Trenton Fossils of Grand Isle," *Vt. State Geologists' Rept. for 1919–20*, pp. 90–101. Gilbert Van Ingen, "The Potsdam Sandstone of the Lake Champlain Basin," *N.Y. State Mus., Bull. 52* (1902), pp. 528–45.

[3] J. F. Kemp and V. F. Marsters, "Trap Dikes of the Champlain Region," *U.S. Geol. Surv., Bull. 107* (1893). H. W. Schimer, "Petrographic Description of the Dikes of Grand Isle," *Vt. State Geologist's Rept. for 1901–2*, pp. 174–83.

of considerable use in locating faults and determining the amount of movement involved. The thicknesses of the formations are indicated by Figure 2.

But it is desirable to discuss the surface of the pre-Cambrian rocks upon which rest the Lower Paleozoic strata, because on the nature of that surface depend certain interpretations of structure. Thus, in the northern part of the Adirondacks, where outcrops are not plentiful, there are apparently many irregularities in the boundary between

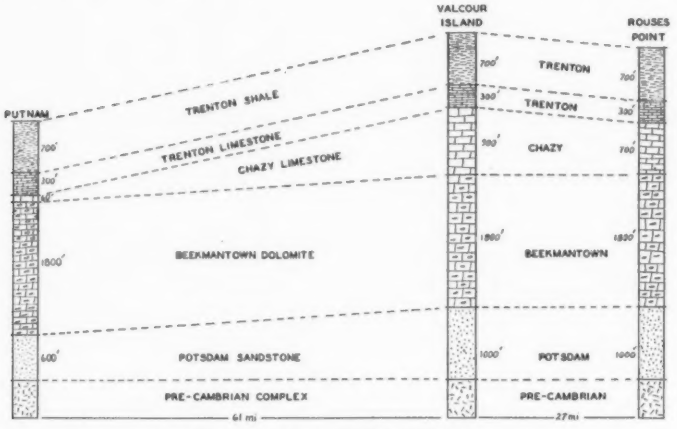


FIG. 2

pre-Cambrian and Paleozoic rocks which admit of two alternative interpretations. The irregularities may be due to inequalities of level of the floor on which the Paleozoic sediments were deposited, or faulting may have lowered or raised blocks so as to make re-entrants or promontories in the line of contact.

To the south of Port Kent, practically all of the pre-Cambrian–Paleozoic boundary line is fault contact, and the re-entrants are obviously due to faulting; but to the north the great development of Pleistocene deposits so masks the structures that such relations cannot be proved or disproved. Consequently, it is necessary to make interpretations on the basis of regional conditions.

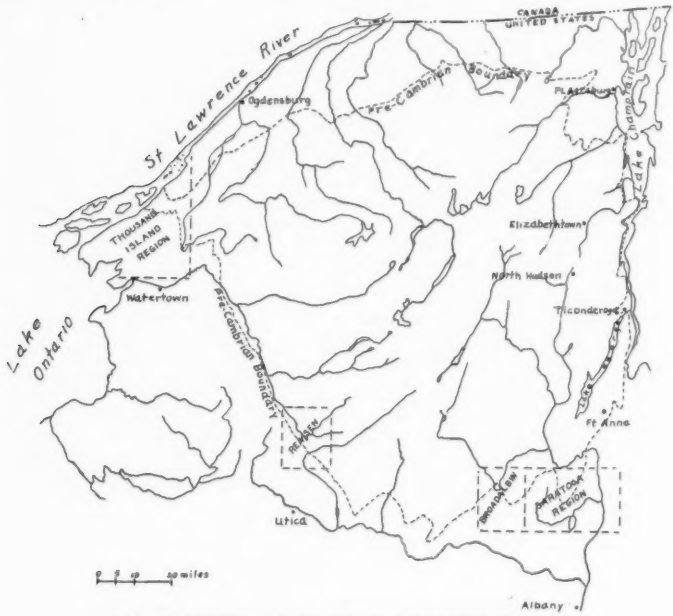
Of the conditions in the northern part of the area it is possible to say only that there seem to be no outliers or inliers of large size and that, therefore, the surface was not very rough. In the eastern and southern parts of the Adirondacks, however, there are many outliers of Paleozoic rocks within the area of pre-Cambrian rocks.

FIG. 3.—Unconformable surface between pre-Cambrian rocks and Potsdam sandstone, Crown Point Center, New York (Ticonderoga sheet).

Kemp[4] has interpreted these as original depressions on the pre-Cambrian floor under the Paleozoic sediments, but, because most of them are located where topographic evidence of faulting is strong and because the rocks in the outliers are lithologically similar to the beds in the Champlain valley where they are normally developed, it is believed that the outliers are due to faulting.

[4] J. F. Kemp, "Pre-Cambrian Topography of the Eastern Adirondacks (abstract)," *Jour. Geol.*, Vol. V (1897), p. 101; "Physiography of the Eastern Adirondacks in the Cambrian and Ordovician Periods," *Bull. Geol. Soc. Amer.*, Vol. VIII (1897), pp. 408–13; "The Cambro-Ordovician Outlier at Wellstown, Hamilton County, New York (abstract)," *Sci.*, Vol. XIII (1901), p. 710, and *N.Y. Acad. of Sci. Annals*, Vol. XIV (1902), pp. 113–15. Rudolf Ruedemann, "Additional Note on Oceanic Currents in the Utica Epoch," *Amer. Geologist*, Vol. XXI (1898), pp. 75–81; "Types of Inliers Observed in New York," *N.Y. State Mus., Bull. 133* (1909), pp. 166–93.

More helpful information upon the nature of the surface of the Basement Complex is contained in the reports of the New York State Museum on the areas at different places along the south and west margins of the Adirondacks. Miller reports that in the Rem-



MAP OF ADIRONDACK REGION SHOWING LOCATION OF PLACES WHERE THE SURFACE BENEATH THE PALEOZOIC ROCKS HAS BEEN STUDIED

FIG. 4

sen[5] and Broadalbin[6] quadrangles the old surface was rather flat, with a relief of 100 feet or less. Cushing and Ruedemann, in their work in the Saratoga and Schuylerville quadrangles,[7] found a relief

[5] W. J. Miller, "Geology of the Remsen Quadrangle, Including Trenton Falls and Vicinity in Oneida and Herkimer Counties," *N.Y. State Mus., Bull. 126* (1909), pp. 35–36.

[6] "Geology of the Broadalbin Quadrangle," *ibid., Bull. 153* (1911), pp. 50–52.

[7] H. P. Cushing and Rudolf Ruedemann, "Geology of Saratoga Springs and Vicinity," *N.Y. State Mus., Bull. 169* (1914), pp. 31–32.

of 75 feet or more, with elevations and depressions not sharp but gentle, as on a surface in old age. The Thousand Islands region[8] is perhaps the most favorable place for studying the nature of this surface, because there the present surfaces on the pre-Cambrian rocks are believed to represent practically the old depositional floor, from which the Potsdam sandstone has lately been removed by erosion. In this locality an abundance of facts shows that the old land had a knob and basin topography with low ridges and valleys trending northeast-southwest. There were few flats and the maximum relief was about 100 feet.

The evidence just considered indicates that south of a line running from about the region of Ticonderoga to the Thousand Islands region the old surface was fairly smooth. The absence of prominent outliers and inliers north of that line shows that no great irregularities existed there. Cushing says of the surface in the northern Adirondack area, "Maximum differences of level of but a few hundred feet are all that are involved, and these comparatively seldom."[9] The conclusion is reached, therefore, that the old land surface had no greater relief than 200 or 300 feet, and that greater differences in elevation now found within small areas on that surface must be due to post-Cambrian faulting.

GENERAL CHARACTERISTICS OF THE FAULTS

The faults of the Lake Champlain lowlands are all, as far as can be determined, of the normal, or gravity, type. The downthrows of those faults trending in a north-south direction are almost all to the east and in consequence the pre-Cambrian surface under the Paleozoic rocks descends by steps to the east. A possible exception is the fault at the shore near Arnold Bay, a little south of west from Vergennes, Vermont, but the strike of this fault changes to northeast-southwest a short distance north of Arnold Bay. The downthrows of the northeast-southwest and other faults may be on either side. There are not many horsts or graben bounded by parallel faults, but, at the intersection of faults, angular blocks may be lower or

[8] H. P. Cushing, H. L. Fairchild, Rudolf Ruedemann, and C. H. Smyth, Jr., "Geology of the Thousand Islands Region," *N.Y. State Mus., Bull. 145* (1910), pp. 54–60.

[9] "Geology of the Northern Adirondack Region," *op. cit.*, p. 419.

higher than the surroundings. It is common for the beds of the downthrown blocks to dip west, toward the upthrown block, the "backward tilting" of Swinnerton.[10] The stratigraphic throws vary greatly in amount, the largest being 4,000 feet, just north of Split Rock Point. Many of those along the pre-Cambrian boundary are about 1,000 feet, as minima. There are many, also, too small to map. G. H. Hudson has discovered and mapped a surprising number of small faults on Valcour Island.[11] In general, the faults are straight, although curving is not uncommon, and a few of them branch.

SYSTEMS OF FAULTS

On Figure 5, which is a map of the region simplified to show the relations of the faults to each other, it is apparent that the faults trend in two main directions, northeast-southwest and north-south. There are some, also, which strike northwest-southeast, and a few which strike east-west. In addition, there are a few small faults with no systematic orientation, which are entirely confined to blocks outlined by the larger faults and which are probably minor fractures formed during the movements of the larger blocks.

The question arises as to whether these sets of faults represent different episodes of faulting. If they do, there is some reason for thinking that the north-south set was formed first. A glance at Figure 5 will show that several of the north-south faults appear to be cut off by those of the northeast-southwest system. This can be seen especially well along the boundary of the Adirondack pre-Cambrian rocks southwest of Willsboro. At no place, however, is a fault of one system seen to be actually offset by one of another system, and the relations are not considered to be definite enough to prove that one system of faults is older than another. On the contrary, there is a situation which suggests that the different systems were developed at the same time. In this case a fault of one system intersects one of another without there being a continuation of either beyond the point of intersection (see the pre-Cambrian boundary north of Ticonderoga and Figure 6). The relations at the points indi-

[10] A. C. Swinnerton, "Block Fault Structure Near Ticonderoga, New York (abstract)," *Bull. Geol. Soc. Amer.*, Vol. XLII (1931), p. 202; "Structural Geology in the Vicinity of Ticonderoga, New York," *Jour. Geol.*, Vol. XL (1932), pp. 402–16.

[11] Unpublished manuscript; "The Fault Systems of the Northern Champlain Valley, New York," *N.Y. State Mus., Bull. 286* (1931), pp. 5–81.

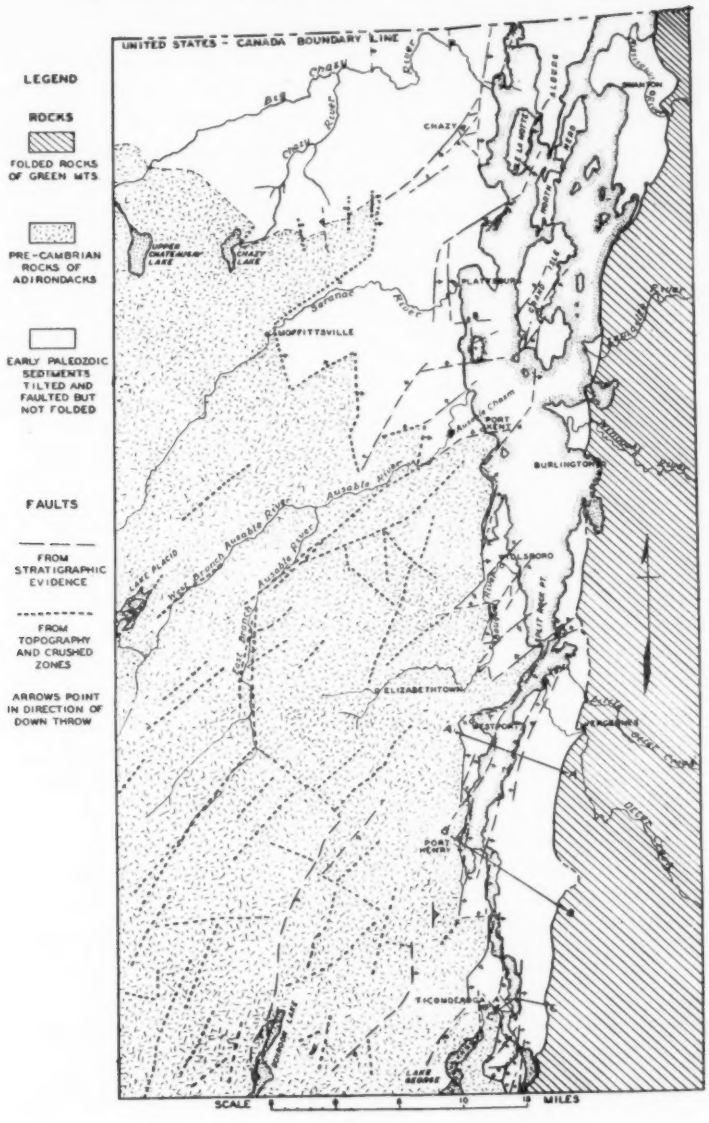

LEGEND
ROCKS
FOLDED ROCKS OF GREEN MTS.
PRE-CAMBRIAN ROCKS OF ADIRONDACKS
EARLY PALEOZOIC SEDIMENTS TILTED AND FAULTED BUT NOT FOLDED
FAULTS
FROM STRATIGRAPHIC EVIDENCE
FROM TOPOGRAPHY AND CRUSHED ZONES
ARROWS POINT IN DIRECTION OF DOWN THROW
UNITED STATES - CANADA BOUNDARY LINE
CHAZY
PLATTSBURG
MOFFITTSVILLE
PORT KENT
BURLINGTON
KEESEVILLE
ELIZABETHTOWN
WESTPORT
PORT HENRY
TICONDEROGA
LAKE GEORGE
SCALE
MILES

FIG. 5

cated show that the mass outlined by the faults moved as a unit. In Figure 6 it is plain that faults with north-south, northeast-southwest, east-west, and northwest-southeast strikes moved simultaneously. The same interpretation is attached to the curving of the faults in the Mohawk Valley from north-south to northeast-southwest and back to north-south again. Cushing and Ruedemann, in their report on the Saratoga region, suggest that the movements along these curving faults relieved tension acting in a direction intermediate to the directions which would produce either north-south or northeast-southwest faults alone.[12] In other words, the tension, instead of producing one set of fractures normal to the direction of tension, produced faults in two directions, the two directions being old lines of weakness in the basement rocks. Similarly for the Lake Champlain region, it is believed that rather simple tensile forces caused complex movements along lines of weakness in the Basement Complex.

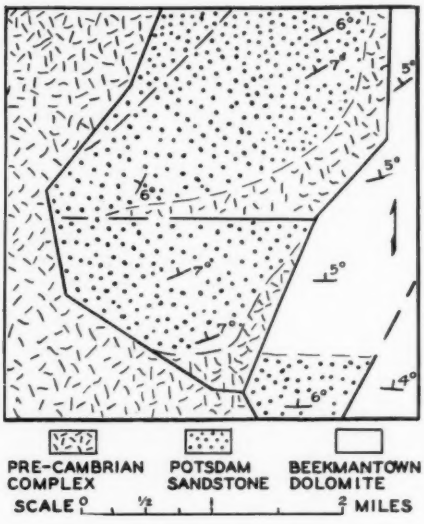


FIG. 6.—Fault blocks south of Ticonderoga

The significance of the foregoing facts is believed to be not that the different systems of faults represent different episodes of faulting, but that there was one period of faulting during which movement was taking place simultaneously in several directions.

DIPS OF THE FAULT PLANES

Previous writings on the faulting in this region contain statements or assumptions that the fault planes are vertical or nearly so, but the facts learned in the present investigation tend to oppose that view.

[12] "Geology of Saratoga Springs and Vicinity," *op. cit.*, p. 64.

Not many of the fault planes can actually be seen, but a few are well exposed. At Treadwells Mill (Plattsburg sheet) on the Saranac River about 3 miles southwest of Plattsburg, New York, there are some very small faults in the Beekmantown formation which strike N. 54° E. and dip 65° SE. Ausable Chasm (Plattsburg sheet) contains a number of faults of unknown, but probably small, displacements in the Potsdam sandstone. A few of them are almost vertical, although most of them have dips of 60°–70°. One of the best exposures of a fault in the whole region is on the shore at the south edge of the village of Essex, New York (Willsboro sheet). There the Trenton shale is thrown against the Chazy limestone by a fault which strikes N. 47° E. and dips 60° SE. On the Bouquet River about $2\frac{1}{2}$ miles west of Essex a fault between the Potsdam sandstone and the pre-Cambrian rocks is seen to have a strike N. 45° E. and a dip 65° SE. At Whallon Bay (Willsboro sheet) just west of Split Rock Point there is a remarkably good exposure of a fault between Trenton limestone and pre-Cambrian rocks. This fault, which is the greatest known in the region, strikes N. 60° E. and dips 55° NW. At Arnold Bay about 6 miles west of Vergennes, Vermont (Port Henry sheet), is a north-south fault with a downthrow to the west. It lies between Trenton shale and Beekmantown limestone, strikes N. 9° E. and dips 45° W. A photograph of a fault at Bluff Point (Plattsburg sheet) was made in 1897 by G. H. Hudson.[13] At the time the writer visited the locality the fault plane was covered with slump material and refuse from nearby camps, so that no measurements could be obtained, but it appears from the photograph that this fault plane must dip only about 65°.

Thus, it is seen that nearly all the fault planes exposed depart considerably from the vertical. The average dip is probably between 60° and 70°. Sixty-five degrees is the figure used in calculating the amount of stretching, except where the dips of the fault planes are known to be different.

FAULTS IN THE ADIRONDACKS

In this investigation faults were mapped only in the areas of Paleozoic rocks, but it is known that the same systems of faults ex-

[13] G. H. Hudson, "Interesting Geologic Features at Champlain Assembly, Cliff Haven," *N.Y. State Mus., Bull. 196* (1917), pp. 149–60, Fig. 5.

tend into the area of the crystalline rocks in the Adirondacks, for, although their detection in the crystalline rocks is not so easy as it is in the sedimentary rocks, the existence of many of them is demonstrated by a variety of evidence. Aside from observed continuations of the faults in the Paleozoic rocks, the evidence consists of four types of facts: namely, long straight boundary lines between formations, crushed zones, topographic expressions of faulting, and outliers of Paleozoic rocks within the area of pre-Cambrian rocks. Those faults within the crystalline area which are shown on Figure 5

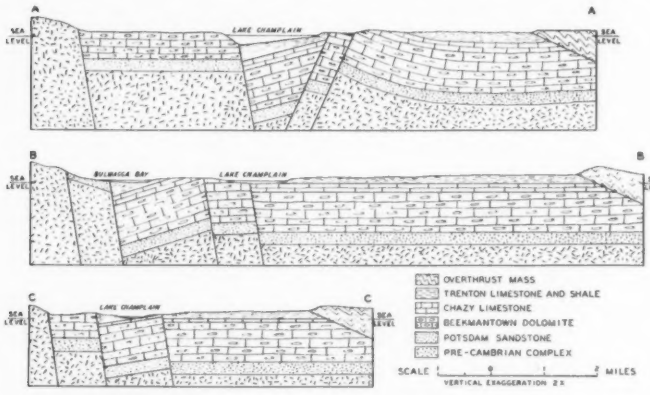


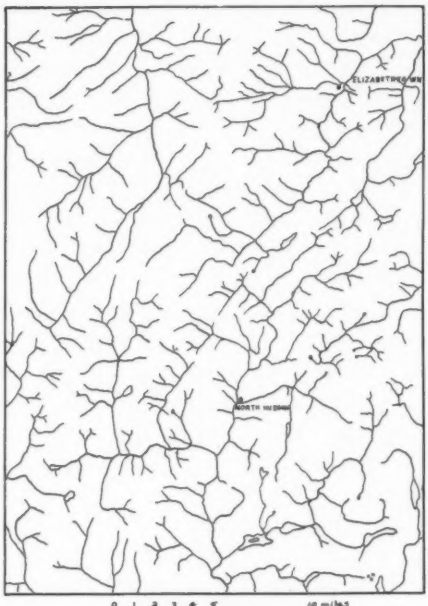
FIG. 7.—Cross sections. For locations see Fig. 5

have been taken mainly from the reports of the New York State Museum.[14] The reader is referred to those reports for the proof of the faults, although Figure 8 is given here to show some of the topographic evidence for faulting in the Adirondacks. Balk says, however, that many of the northeast-southwest lines in the topography are due to closely spaced regional tension joints in the anorthosite.[15]

[14] Kemp and Ruedemann, "Geology of the Elizabethtown and Port Henry Quadrangles," *op. cit.;* J. F. Kemp, "Geology of the Mount Marcy Quadrangles," *N.Y. State Mus., Bulls. 229 and 230* (1921); J. F. Kemp and H. L. Alling, "Geology of the Ausable Quadrangle," *N.Y. State Mus., Bull. 261* (1925); W. J. Miller, "Geology of the Schroon Lake Quadrangle," *N.Y. State Mus., Bulls. 213 and 214* (1919); I. H. Ogilvie, "Geology of the Paradox Lake Quadrangle," *N.Y. State Mus., Bull. 96* (1905).

[15] Robert Balk, "Structural Geology of the Adirondack Anorthosite," *Mineralogische und Petrographische Mitteilungen,* Band 41 (1931), p. 419.

It is of considerable importance that these faults in the Adirondacks belong to the same systems as do those in the lowlands. There can hardly be any doubt that they are part of the same deformation.



ADIRONDACK DRAINAGE
(From New York State Geological Map)

FIG. 8

MOHAWK VALLEY AND ASSOCIATED FAULTS

That the fault structure of the Lake Champlain region extends south of the limits of mapping done by the author is well shown by the early work of Vanuxem, Hall, and Darton and the more recent work of Cushing and Ruedemann, Miller, and others.[16] The full ex-

[16] Lardner Vanuxem, *Natural History of New York*, Part III, Third Geological District (1842), pp. 203–11. C. E. Hall, "Field Notes on the Geology of the Mohawk Valley," *N.Y. State Geologist, Ann. Rept. 5* (1886), pp. 8–10, and *Ann. Rept. 14* (1895), pp. 54–56. N. H. Darton, "Geology of the Mohawk Valley," *Forty-seventh Ann. Rept., N.Y. State Mus.* (1894), pp. 601–25; "Preliminary Description of the Faulted Region

tent and characteristics of the faulting however, have not been ascertained.

The Mohawk Valley faults are mainly between Schenectady and Little Falls. Most of them have rather small displacements where they cross the Mohawk River, the stratigraphic throws given being 200–310 feet. To the north, as the faults run into the Adirondacks, the displacements become greater. They strike mainly northeast-southwest and most of them have downthrows to the east. They curve considerably, varying between northeast-southwest and north-south, and branch at several places. The fault planes exposed are almost vertical.

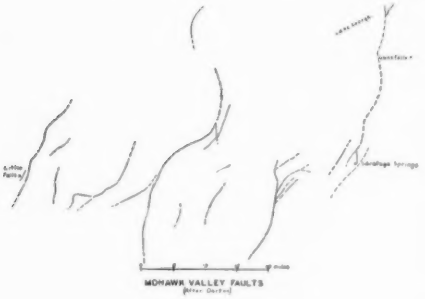


FIG. 9

These faults do not extend far south of the Mohawk Valley. The northern limit of the faulting has not been determined. Most of the faults were traced by Darton north to the crystalline rocks of the Adirondacks and one was traced by him into the crystalline rocks as far as the outlier at Wells.[17] At the east end of his map there is a long fault which runs northeast-southwest through the Saratoga region as far as the south end of Lake George. In fact, Darton believed the lake basin to have been excavated out of Paleozoic rocks on a fault block.[18] The writer has mapped faults into the north end of Lake George; so it seems that there must be a connection between the Mohawk Valley faults and the Lake Champlain faults.

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of Herkimer, Fulton, Montgomery, and Saratoga Counties," *Forty-eighth Ann. Rept., N.Y. State Mus.* (1894), pp. 30–56. Cushing and Ruedemann, "Geology of Saratoga Springs and Vicinity," *op. cit.* Miller, "Geology of the Remsen Quadrangle, etc.," *op. cit.;* "Trough Faulting in the Southern Adirondacks," *Sci.*, Vol. XXXII (1910), pp. 95–96; "Geology of the Broadalbin Quadrangle," *op. cit.*

[17] Darton, "Preliminary Description of the Faulted Region of Herkimer, etc.," *op. cit.*, p. 47.

[18] *Ibid.*, p. 52.

Typical Lake Champlain structure is present in the Saratoga region. Cushing and Ruedemann[19] describe the structure west of the overthrusts as characterized by normal faults with few folds. A general northeast-southwest strike is present, but there is a noticeable tendency of the faults to curve from northeast-southwest to north-south and back to northeast-southwest. Cushing and Ruedemann suggest that this may be due to the fact that tension toward the southeast was relieved along two previously existing directions of weakness. The downthrows are mainly to the east, although a few are to the west. The blocks are tilted slightly to the west and only a few horsts and graben are present. The fault planes are reported to be vertical. The stratigraphic throws are rather great, as may be seen from the following estimates: Hoffman's Ferry fault, 1,300–1,600 feet; West Galway fault, 250–300 feet; McGregor fault, over 1,400 feet; Woodlawn Park fault, 300 feet. It is interesting to note here that none of these normal faults has been found to cut the overthrust faults east of the normal-fault area and none has been found in the overthrust masses. The writers of the report believe this to mean that the normal faulting was earlier than the overthrust faulting.[20]

Lake George, as stated before in this paper, forms a connecting link between the faulted area studied by the writer and the faulted area of the Mohawk Valley. Faults have been traced to both ends of Lake George. Certain of them obviously run some distance in the lake. Kemp,[21] in discussing the physiography of the lake, concluded that faulting has been very important in the formation of the basin. The connection between the two areas would, without doubt, be more apparent if all the intervening region were mapped in detail, but even with the evidence at hand it appears certain that the faulting of the Mohawk Valley is the same as that in the Lake Champlain region.

[19] Cushing and Ruedemann, "Geology of Saratoga Springs and Vicinity," *op. cit.* pp. 53–65.

[20] *Ibid.*, p. 145.

[21] "Notes on the Physiography of Lake George (abstract)," *Amer. Geologist*, Vol. XXVIII (1901), pp. 331–32; *Sci.*, Vol. XIV (1901), p. 774; and *N.Y. Acad. Sci. Annals*, Vol. XIV (1902), pp. 141–42.

WESTERNMOST OVERTHRUST

The westernmost overthrust of the Green Mountains was chosen as the eastern limit to the area of this study, because it marks the boundary between the lowland area of tensional structures and the

FIG. 10.—Overthrust of Chazy limestone on crumpled Trenton shale, Lime Rock Point (St. Albans sheet).

mountain region of compressional structures. The boundary drawn on the maps does not represent a single overthrust fault, for in certain places, where small slices were found underneath the larger overthrusts, the line was drawn along the underlying minor fault. At other points the boundary is along the main overthrusts. No attempt has been made to show the complexity of the overthrusts nor

to indicate on the maps what rocks compose the overthrust block, because the overthrusts were studied only to locate the boundary of the lowland area and to determine whether or not they were cut by the normal faults.

A satisfactory answer to the question of whether the normal faults cut the overthrusts was not obtained. There are localities, such as that east of Split Rock Point, where normal faults of considerable size trend directly toward the overthrusts and where it would seem that the true relations could be determined, but at these places the cover of Pleistocene material happens to be so complete that the relations cannot be determined definitely.

The evidence, as far as the writer knows, seems to indicate that the normal faults are older than the overthrusts and do not cut them. This belief is supported by the fact that the overthrusts are definitely not cut by normal faults over certain rather great distances, as in the northern half of the Milton quadrangle and south of Vergennes, Vermont (Port Henry sheet), where the overthrust is marked by a continuous limestone ridge 8 or 9 miles long. These distances seem too great to have escaped faulting by mere chance. The general straightness of the overthrust line as compared, for example, with the pre-Cambrian boundary line of the Adirondacks, where faults are known to be present, also favors this idea. Offsets of the overthrust faults by normal faults might exist without detection in a few places, however, such as east of Split Rock Point, at Vergennes, and southwest of Snake Mountain (Port Henry sheet), but there is no strong evidence for offsets in those localities.

CRUSTAL STRETCHING

Figure 11 is an attempt to portray the crustal stretching involved in the Lake Champlain faulting. The vectors show the amount and direction of stretching and, consequently, indicate something of the nature of the forces which caused it. The basis for calculation is that the heave of the faults is the amount of stretching. The fault planes have been assumed to dip 65°, except where actually known to be different. This assumption appears to be justified by measurements of fault planes exposed at various places in the region. It has been assumed, also, that all the movements have been dip-slip movements,

because the few striations which have been observed were nearly in the directions of the dips of the fault planes. Each vector was con-

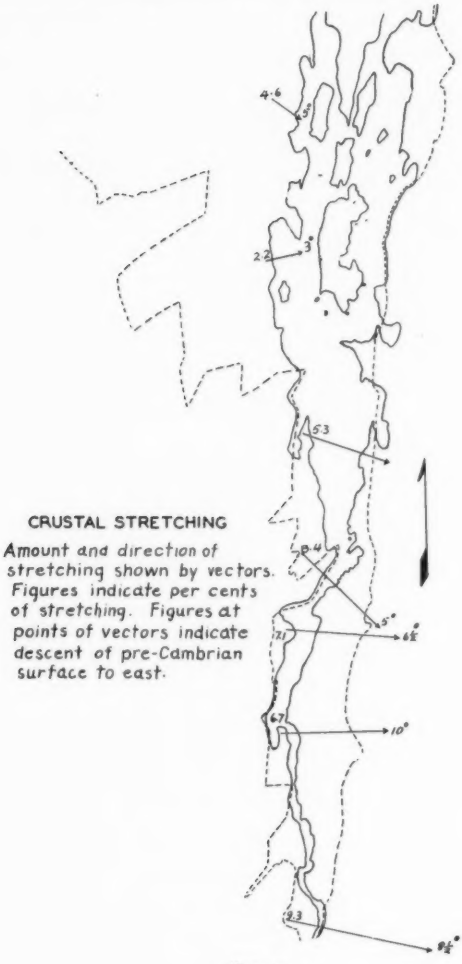


FIG. 11

structed for the faults in a narrow zone across the lowland area at localities where the faults are fairly well known. Except for the two

northernmost vectors the zones extend from the pre-Cambrian rocks of the Adirondacks to the easternmost point at which the structure appears to be rather definitely known, and do not include the shale areas just west of the overthrusts where faults cannot be detected. The amount of stretching, in percentages, is indicated by the figure at the base of each vector.

The surface at the base of the Paleozoic series is known to descend to the east by steps. The amount of that descent is indicated by the figures at the points of the vectors. The figures represent the dip of a line from the pre-Cambrian rocks along the Adirondack boundary to the base of the Potsdam sandstone at the easternmost point where the elevation of that horizon can be determined.

These reckonings involve several uncertainties and, consequently, are only approximate. Some of the matters of uncertainty are: the dip of the fault planes, the direction of movement on the fault planes, the possibility that the thickness of the lower rocks may change to the east, the possibility of undetected faults, and the possibility of error in the calculation of the stratigraphic throws.

Because of the uncertainties involved, these calculations cannot be considered accurate. They mean little more than that there were tensional forces from the east and southeast, that the forces were greater or their results more pronounced in the southern part of the area, that the pull was downward to the east, and that the downward component of the forces was greater or its effect more evident in the southern part of the area.

ASSOCIATION WITH GREEN MOUNTAINS AND ADIRONDACK MOUNTAINS

Geographically the Lake Champlain normal faults are associated with the Green Mountains and the Adirondack Mountains. These faults are found only in a broad zone west of the Green Mountains, and the intensity of the faulting appears to be greatest near the Green Mountains. On the other hand, the normal fault structure is not present in those mountains nor has it been found throughout their western border zone. Apparently, it does not extend far south of the Mohawk Valley, and, although it reaches Canada on the north, it seems to be diminishing in that direction (see Fig. 11). In fact, the greatest development of the normal fault structure seems

to be in the vicinity of the Adirondack Mountains. In the opinion of the writer, the association with the Adirondacks is secondary to the main association with the Green Mountains. This seems true, because the pattern is not the kind to have been produced by the Adirondack doming movement, for it is neither radial nor concentric as might be expected if caused by doming. Neither is the distribution of the faults that which would be expected if they were primarily related to the almost circular Adirondack mass, for they are almost all in the eastern half of the dome[22] and seem to follow the Green Mountain front. Furthermore, the direction of the faults and of the stretching (see Figs. 5 and 11) does not vary from place to place about the margin of the Adirondacks as it should if it were caused by the same forces that raised the Adirondacks.

It is suggested that these relations are the result of forces which were present all along the border of the Green Mountains and associated ranges, but which were noticeably active only in the vicinity of the upstanding Adirondack area.

DATE OF THE NORMAL FAULTING

The foregoing discussion of the systems of faults has shown that the normal faults of this region were produced simultaneously. The faulting is, therefore, considered as one deformational event, whether it occurred in a short space of time or whether it extended over a great length of time. There is not a question of dating the different systems.

STRATIGRAPHIC EVIDENCE

The youngest rock affected by this normal faulting is the Trenton shale. The oldest rocks known to be unaffected are Pleistocene sediments. The relation of the dikes to the faults is unknown, and the age of the dikes is not known any more closely than that of the faults. The stratigraphic relations obviously do not furnish much information about the time of faulting, for they show only that it was post-Trenton and pre-Pleistocene.

STRUCTURAL EVIDENCE

The structural relations give closer limits to the date of faulting. As far as has been determined in this region the normal faults do not

[22] Cushing, "Geology of the Northern Adirondack Region," *op. cit.*, p. 406.

cut the overthrust faults. There seems to be no reason why the normal faulting should not extend into the area of the overthrusts, for the normal faults are found immediately west of it. They are not so abundantly evident near the overthrusts, however, but the difficulty of detecting them in the shale areas along the overthrusts probably accounts for the scarcity of known faults there. This does not explain the absence of the normal faults in the overthrust blocks nor the lack of offsets of the overthrust planes, because in certain areas, at least, the overthrust masses present good mapping conditions. It is believed that the absence of the normal faults in the overthrust area is due to their having been covered by the overthrust masses. If so, the normal faults are older and the time of overthrusting is the later limit to the time of normal faulting. Cushing and Ruedemann incline toward this view for the Saratoga region.[23]

TIME OF THE OVERTHRUSTING

With the time of normal faulting placed between the Trenton epoch and the time of overthrusting, the date of the overthrusting is next to be considered. This compression has been assigned to three different deformations, the Taconic Disturbance, the Acadian Disturbance, and the Appalachian Revolution.

In the Lake Champlain region the stratigraphic relations prove only that the deformation was post-Trenton and pre-Pleistocene in age. In recent years considerable doubt has been cast on the reality of the Ordovician disturbance, particularly its effect in Vermont.[24] Because the folds and overthrusts of the Green Mountains connect with and have the same characteristics as those in the Appalachians, some have assigned the Green Mountain deformation to the Appalachian Revolution.[25] Most recently, however, Schuchert has reviewed the field evidence and concluded that the "Taconian" disturbance at the end of the Ordovician period was the main deformation affecting the Green Mountains.[26] The Acadian and Appalachian

[23] "Geology of Saratoga Springs and Vicinity," *op. cit.*, p. 145.

[24] T. H. Clark, "Review of the Evidence for a Taconic Revolution," *Proceedings Boston Soc. Natural Hist.*, Vol. XXXVI (1921), pp. 135–63.

[25] Arthur Keith, "Cambrian Succesion of Northwestern Vermont," *Amer. Jour. Sci.*, Vol. V (1923), pp. 97–139, and *Vt. State Geologist's Rept. for 1923–24*, pp. 105–36.

[26] Charles Schuchert, "Orogenic Times of the Northern Appalachians," *Bull. Geol. Soc. Amer.*, Vol. XLI (1930), pp. 701–24.

deformations were important, but they affected areas farther to the east.

Conclusion.—It is concluded that the Lake Champlain normal faults were developed after the Trenton epoch and before the Taconic Disturbance, or in late Ordovician time.

ASSOCIATION OF NORMAL FAULTS WITH OVERTHRUSTS

If the association of the normal faults of the Lake Champlain region with the Green Mountain structure is due to causal relation of the two structures, there should be similar associations of normal faults with overthrust faults elsewhere. A few random examples of such relations will be cited in the following pages.

GREEN MOUNTAINS AND ASSOCIATED RANGES

It might seem that normal faults should be found all along the west side of the Green Mountains and associated ranges. As a matter of fact, no great development of them is seen south of the Mohawk Valley, though scattered faults exist in the Catskills and southward. How far into Canada the structure extends is not known, but the intensity of the faulting appears to diminish north of the Adirondacks. Thus, it seems, although the pattern does not appear to have been produced by the Adirondack doming, that this faulting is confined to the vicinity of the Adirondacks. In a foregoing section it was suggested that the tension postulated along most of the length of the present mountain border was relieved by barely noticeable adjustments over wide areas, whereas adjacent to the more rigid and upstanding Adirondack mass it was concentrated to produce faults.

JURA TABLELAND

An interesting example of normal faulting associated with compressional structures exists in the Jura Tableland north of the Jura Mountains.[27] About eighty important normal faults are known in this region, ranging up to 1,500 meters in stratigraphic throw. Most of them have a north-south strike, but a few are east-west. The fault planes dip from 80° to 85°. Horst and graben structure is characteristic. These faults were formed before the time of the last thrust movements in the Juras.

[27] L. W. Collet, *Structure of the Alps* (London: Edward Arnold & Co., 1927), pp. 134–38; A. Heim, *Geologie der Schweiz*, Band I (1919), pp. 553–65.

WESTERN UNITED STATES

In the western regions of the United States are several examples of normal faults associated with overthrusts. Certain of the normal faults in the foothills in Wyoming and Montana appear to have been formed during the compression of the Rocky Mountain beds.[28] They are, for the most part, fractures in dome structures. They differ, also, from the Lake Champlain faults, as they were formed during the compression rather than before it.

In the Lewis and Livingston ranges of Montana there are said to be normal faults west (back) of the overthrusts.[29] These, however, are younger than the overthrusts and cut them.

The normal faults which are known to be present near the Absaroka-Owl Creek Mountain front and in the margin of the Bighorn basin were formed soon after the development of the Heart Mountain overthrust.[30] They have not been studied in detail.

In southwestern Wyoming there is a belt of normal faults just east of the Absaroka overthrust in the overridden mass.[31] The normal faults have not been mapped and the relations to the overthrusts are not well known, but they have been ascribed to settling after the overthrusting.

Normal faults are associated with the Bannock overthrust fault in Idaho and in northern Utah. Both types of faults have been studied by Mansfield[32] who states that the normal faults are relaxa-

[28] T. A. Link, "The Origin and Significance of 'Epi-Anticlinal' Faults as Revealed by Experiments," *Bull. Amer. Assoc. Pet. Geologists*, Vol. XI (1927), pp. 853–67.

[29] R. G. McConnel, "Report on the Geological Structure of a Portion of the Rocky Mountains, Accompanied by a Section Measured near the Fifty-first Parallel," *Can. Geol. Surv.* (1887); Bailey Willis, "Stratigraphy and Structure of the Lewis and Livingston Ranges, Montana," *Bull. Geol. Soc. Amer.*, Vol. XIII (1902), pp. 305–52.

[30] D. F. Hewett, "The Heart Mountain Overthrust, Wyoming," *Jour. Geol.*, Vol. XXVIII (1920), pp. 537–57.

[31] A. C. Veatch, "Geography and Geology of a Portion of Southwestern Wyoming, with Special Reference to Coal and Oil," *U.S. Geol. Surv., Prof. Paper 56* (1907).

[32] G. R. Mansfield, "Types of Rocky Mountain Structure in Southeastern Idaho," *Jour. Geol.*, Vol. XXIX (1921), pp. 444–68; "Structure of the Rocky Mountains in Idaho and Montana," *Bull. Geol. Soc. Amer.*, Vol. XXXIV (1923), pp. 263–84; "Geography, Geology, and Mineral Resources of Part of Southeastern Idaho," *U.S. Geol. Surv., Prof. Paper 152* (1927). R. W. Richards and G. R. Mansfield, "The Bannock Overthrust, a Major Fault in Southeastern Idaho and Northeastern Utah," *Jour. Geol.*, Vol. XX (1912), pp. 681–709.

tional features which resulted from the overthrusting. Some of them appear to be satisfactorily explained that way, but others, according to his diagrams and cross sections, appear to be older than the overthrusts.

The Wasatch Mountains in Utah contain overthrusts with younger normal faults associated.[33] These faults are also thought to be due to relaxation after overthrusting.

The Spring Mountain Range in southern Nevada furnishes another example of overthrusting followed by normal faulting.[34]

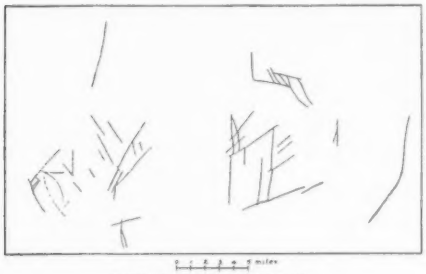


FIG. 12

The Muddy Mountains region of Nevada, recently described by Longwell, has a marked development of overthrusting and normal faulting.[35] Many of the normal faults are younger than the overthrusts, but some are older. Longwell thinks that the older ones may have been produced by relaxation after an earlier overthrusting. It is interesting and perhaps significant that the pattern of the Muddy Mountains faults is very similar to that of the Lake Champlain region.

This review of mountain border normal faults, brief and incom-

[33] G. F. Loughlin, "Reconnaissance in the Southern Wasatch Mountains, Utah," *Jour. Geol.*, Vol. XXI (1913), pp. 436–52; Eliot Blackwelder, "New Light on the Geology of the Wasatch Mountains, Utah," *Bull. Geol. Soc. Amer.*, Vol. XXI (1910), pp. 517–42; Hyrum Schneider, "A Discussion of Certain Geologic Features of the Wasatch Mountains," *Jour. Geol.*, Vol. XXXIII (1925), pp. 28–48; F. B. Stillman, "A Reconnaissance of the Wasatch Front between Alpine and American Fork Canyons, Utah," *Jour. Geol.*, Vol. XXXVI (1928), pp. 44–55.

[34] D. F. Hewett, "Structure of the Spring Mountain Range, Southern Nevada (abstract)," *Bull. Geol. Soc. Amer.*, Vol. XXXIV (1923), p. 89.

[35] C. R. Longwell, "The Muddy Mountain Overthrust in Southeastern Nevada," *Jour. Geol.*, Vol. XXX (1922), pp. 63–73; "Structural Studies in Southern Nevada and Western Arizona," *Bull. Geol. Soc. Amer.*, Vol. XXXVII (1926), pp. 551–84; "Geology of the Muddy Mountains, Nevada," *U.S. Geol. Surv., Bull. 798* (1928).

plete as it may be, suffices to show that an association of normal faults with overthrusts is common.

SUMMARY OF FACTS

Thus far it has been the aim of the writer to give the characteristics, relations, and associations of the Lake Champlain faults, and of other associated faults, for the purpose of determining the origin of the structure. There now follows a summary of the most significant facts which must be explained and from which a theory of origin may be derived.

1. The structure was produced by tensional forces acting with a downward pull toward the east or southeast.

2. The faulting of the different systems occurred simultaneously and, as far as has been determined, it all took place before the overthrusting of the Green Mountains.

3. There is a definite relation of the normal-fault structure to the Green Mountain structure and a somewhat secondary relation to the Adirondack structure.

4. Normal faults commonly occur in association with mountain structures.

POSSIBLE EXPLANATIONS

FRACTURE DURING THE DOMING OF THE ADIRONDACKS

The association of the Lake Champlain faults with the Adirondacks might suggest that the same forces produced both. This has been shown in a foregoing section to be unlikely, however.

COLLAPSE STRUCTURE

Certain areas of complex normal faulting are believed to have resulted from the collapse of the crust during the withdrawal of large quantities of magma from beneath. Areas of this kind are typified by some of the mining districts of the Southwest. That normal faulting would result from such a cause is reasonable, and the association of the faults with the lavas is very definite in certain of these places. Probably the first faults to develop in such a process would follow previously formed joint planes. Later ones would develop partly along joints, but also in other directions according to the stresses

placed on the different blocks during the settling. Faults of this type are known at Tonopah, Nevada;[36] Ray, Arizona;[37] and Bullfrog, Nevada.[38] It is evident in comparing the patterns of these faulted areas that considerable differences exist between the examples of this type of structure. The Tonopah faults are noticeable for the right-angle pattern, the distinct blocks, the straight faults,

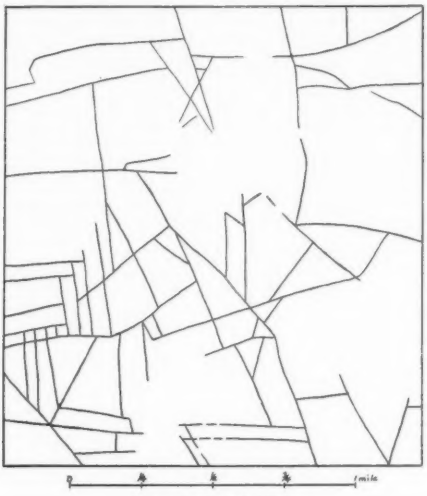


FIG. 13

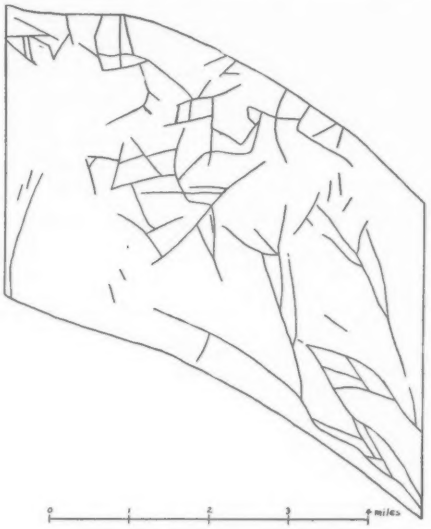
and the existence of long faults cutting many short ones. The Ray district is characterized by the irregularity of the blocks, the presence of several long branching faults, and the curvature of many of the faults. The Bullfrog pattern is marked by a great development of branching and by curvature of the faults in a way to make a sort of net structure. These differences are probably due to differences of the arrangements of the joints in the rocks before faulting, the man-

[36] J. E. Spurr, "Geology of the Tonopah Mining District, Nevada," *U.S. Geol. Surv., Prof. Paper 42* (1905).

[37] F. L. Ransome, "Ray Folio," *U.S. Geol. Surv. Folio 217* (1923).

[38] F. L. Ransome, W. H. Emmons, and G. H. Garrey, "Geology and Ore Deposits of the Bullfrog District, Nevada," *U.S. Geol. Surv., Bull. 407* (1910).

ner of fracture of the rocks, the shape of the magma chamber, the course of events during the extrusion, the distribution of the extrusive rocks, and perhaps to other factors. In spite of all these differences in the patterns, there are certain common characters which distinguish them as a type. The great irregularity and complexity of

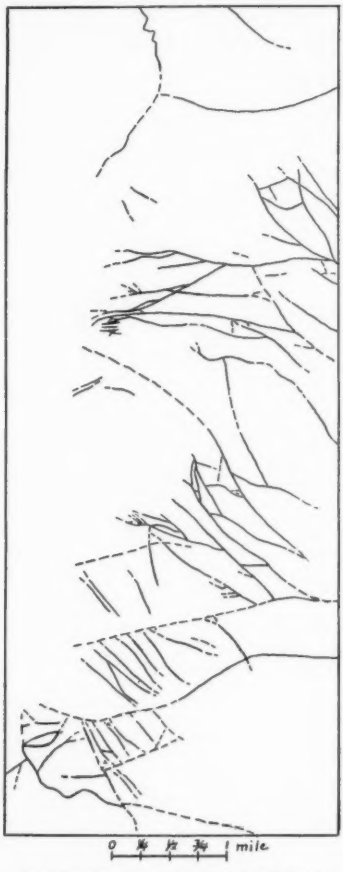


RAY FAULT PATTERN
(After Ransome)

FIG. 14

the faulting are especially apparent. Separate blocks are common, perhaps more common than they appear to be on these maps which are necessarily not complete. Small faults within the blocks are characteristic. Systems of faults are poorly developed. In general, the structure is that of a jumble of broken blocks. It differs from the Lake Champlain structure by a greater complexity, a greater irregularity, and a greater development of separate blocks. The absence of any large amount of extrusive material from the Lake Champlain region and the distance from large-scale igneous activity of any kind also seem to indicate that the Lake Champlain normal faults did

not result from the movement of magmas beneath. It must be admitted, however, that there is not sufficient evidence to show with certainty that magmatic movements had no part in the formation of these faults, for there may have been movements of which we have little or no direct evidence.



BULLFROG FAULT PATTERN
(After Ransome)

FIG. 15

EPI-ANTICLINAL FAULTS

The explanation proposed by Link for certain normal faults in dome structures east of the Rocky Mountain front and for the flaws in the Jura Mountains might be applied here.[39] The idea is, briefly, that tension develops on the outside of an advancing lobe or salient. The tension is in the direction of curvature of the salient, and normal faults develop at right angles to the axes of the folds in the advancing salient. The faults he discusses are mainly within the area of folds, but it seems possible that the stretching would extend out into the unfolded area. The normal faults in the unfolded rocks of the Jura Tableland north of the folded Juras may have been produced in that way. The Lake Champlain faults have about the same position relative to the salient of the Green Mountain folds as the Jura Tableland faults have to the Jura Mountain system. There are, however, certain objections to such an

[39] Link, *op. cit.*; "En Échelon Folds and Arcuate Mountains," *Jour. Geol.*, Vol. XXXVI (1928), pp. 526–38.

explanation for the Lake Champlain faults. The Lake Champlain faults are not normal to the axes of the Green Mountain folds. More important is the fact that the stretching indicated by the Lake Champlain faults is directly opposite to the direction of the thrust forces, and not concentric with the margin of the advancing lobe.

RELAXATION AFTER OVERTHRUSTING

It has been stated in the foregoing pages that many of the normal fault structures associated with overthrust faults and folds are later than the overthrusting and folding. For such normal faults, the common explanation—relaxation after overthrusting—seems reasonable. According to that explanation the loading of certain areas with the overthrust masses and the concentration through folding and overthrusting of the geosynclinal material into a narrower zone cause the newly deformed belts and their borderlands to settle under the increased weight.[40]

For the Lake Champlain faults and others which were formed before the overthrusting, the theory of relaxation after overthrusting is not applicable, unless an earlier time of overthrusting of which there is no evidence is postulated.

TENSION IN THE GEOSYNCLINE

The Lake Champlain faults appear to be as definitely related to mountain structure as are those formed later than the overthrusting, and it is believed that they, too, result from some part of the process of mountain making. That part of the process which precedes the compression is the geosynclinal stage. It is the opinion of the author that these faults were caused by tensional forces which existed between the upstanding Adirondacks and the sinking geosyncline during the geosynclinal stage.

One of the features of the geosynclinal stage is the loading of a narrow zone of the earth's crust with a great mass of sediments. As has been pointed out before, however, the depression of a geosyncline cannot be a simple process of isostatic sinking under the ac-

[40] G. H. Chadwick, "Hypothesis for the Relation of Normal and Thrust-Faults in Eastern New York (abstract)," *Bull. Geol. Soc. Amer.*, Vol. XXVIII (1917), pp. 160–61; J. B. Woodworth, "Relations of Fault-block Mountains to Folded Chains (abstract)," *Bull. Geol. Soc. Amer.*, Vol. XXXI (1920), pp. 115–16; and "Contributions to the Study of Mountain-building" (ed. A. C. Swinnerton), *Amer. Jour. Sci.*, Vol. XXIII (1932), pp. 155–71.

cumulating sedimentary load, for the volume of displaced material at the base of the isostatic column is necessarily less than the volume of the sediments and the sediments will pile up more than they will depress.[41] That still leaves the possibility that, although other forces operate to produce the geosynclinal depression, the weight of the sediments is an added factor of some importance. Lawson says: "The sediment is a load imposed upon the earth's crust and, when the deposit becomes thick enough and the area of the geosyncline is large enough, this load depresses the crust as compensation proceeds, thereby contributing to the development of the geosyncline. The initiation of a geosyncline, however, must be due to some other cause than loading alone."[42] Lawson discusses several typical situations and shows that where areas subject to loading are adjacent to rising areas there must develop shear strains between the rising and the sinking areas which result in flexure or normal faults. The relations in the Lake Champlain region approximate, although they are not exactly the same as,those postulated by Lawson. The normal faults appear to have been formed between the rising Adirondack area and the sinking geosyncline. The distribution and direction of the tensions, as indicated by the faults, are according to expectations; the time of faulting is in agreement with the theory; and the relation of the faults to the Adirondacks and the Green Mountains is explained. The directions of the faults were probably controlled in part by pre-existing lines of weakness. It is not certain, however, that the sedimentary loading would develop enough tension to cause the faults, or that the geosynclines are not downwarps caused by compressive forces that would prevent the development of any tension. Indeed, it would appear that the sedimentary load could cause tension only if the geosyncline were in a state of little or no compression, and that raises the question of how geosynclines are produced.

W. H. Bucher has advanced the idea that geosynclines are produced by tension and that the geosynclinal stage is characterized by a state of tension. That might explain these faults, with or with-

[41] A. C. Lawson, "The Geological Implications of the Doctrine of Isostasy;" *Bull. National Research Council*, Vol. VIII (1924), pp. 3–22; R. T. Chamberlin, "Isostasy from the Geological Point of View," *Jour. Geol.*, Vol. XXXIX (1931), p. 11.

[42] Lawson, *op. cit.*, p. 17.

out the added effect of the sedimentary load. A full statement of this theory has not been made, but the idea of particular significance here is that ". . . . 'mobile belts,' such as the Appalachian and Alpine systems . . . . owe their origin to tensional crustal stresses which create the geosynclines while later compressive stresses throw them into folds."[43] Thus, he postulates tension during the geosynclinal stage, but does not attribute it to the weight of sediments. If geosynclines originate from such tensional stresses and develop in a state of tension, then normal faults might well be produced along the margins of the geosynclines, especially near areas which tend to rise rather than to sink with the geosynclinal areas. Full consideration of this theory cannot be made until a complete statement of it has appeared.

[43] W. H. Bucher, "The Mobile Belts of the Earth (abstract)," *Jour. Wash. Acad. Sci.*, Vol. XXI (1931), pp. 489–91; "The Pattern of the Earth's Mobile Belts," *Jour. Geol.*, Vol. XXXII (1924), pp. 265–90.

THE WANING OF THE LAST ICE SHEET IN CENTRAL MASSACHUSETTS

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ABSTRACT

The modified and unmodified drift deposits of central Massachusetts furnish evidence which seems to indicate that after the ice cover had vanished from the higher lands, long, narrow tongues of ice, many miles in length, still occupied the deeper valleys, and that the last stagnant remnants of these ice masses did not disappear from the valleys until after deposition of modified drift had ceased. The esker ridges, kames, kame terraces, and sand plains were deposited, either against the margin of these masses of ice or in crevasses which dissected them. Limited movements of the ice tongues occurred after some of the terraces and associated structures were formed.

INTRODUCTION

One of the unsettled problems confronting glacial geologists at the present time is the question concerning the conditions which prevailed in New England during the waning of the last, or Wisconsin, ice sheet. One extreme opinion is held by Antevs, Reeds, and many others, who believe that during the retreat the ice sheet maintained a prominent front which gradually withdrew toward the north. Antevs[1] states his view in this sentence: "Accordingly, the Pleistocene ice-sheets underwent appreciable depletion over relatively narrow margins, and normally as they waned preserved a marked front comparable with that of modern ice-sheets." Reeds[2] states his idea in this way:

> The direction of ice retreat from the terminal moraine on Staten Island and Long Island was northward up the Hackensack, Hudson, Connecticut, and smaller river valleys. As the ice border, which extended in a general east-west direction, retreated northward at a rate somewhat less than 100 feet per year, clays were deposited in marginal glacial lakes in the separate river basins.

The other extreme view is held by Flint[3] who maintains that: ". . . . ice must have remained over the present site of Long Island Sound

[1] Ernst Antevs, "Late-Glacial Correlations and Ice Recessions in Manitoba," *Can. Dept. Mines, Geol. Surv. Mem. 168* (1931), p. 28.

[2] C. A. Reeds, "Weather and Glaciation," *Bull. Geol. Soc. Amer.*, Vol. XL (1929), p. 624.

[3] R. F. Flint, "The Glacial Geology of Connecticut," *State Geol. and Nat. Hist. Surv. Bull. No. 47* (1930), p. 67.

Since the manuscript of this paper was submitted for publication, Flint has pub-

and in the lower valley of the Connecticut until the ice had vanished entirely from northern New Hampshire and Vermont." An intermediate position is taken by Andersen[4] who maintains that there were two major stages in the waning ice front; during the first stage the ice was moving forward and able to push up moraines and other deposits under the ice; during the second major stage the ice practically ceased to move and became stagnant or "dead." The ice of this second or stagnant stage occupied a zone around the margin of the moving ice, sometimes very narrow and sometimes very broad. He finds evidence that in Denmark this stagnant zone attained a width of 50 miles.

Conditions which prevailed during the closing stages of the last ice sheet are recorded in the drift deposits, either modified or unmodified, which were left by the retreating ice. In the Nashua River valley in central Massachusetts there is an extensive series of these drift deposits consisting of kames, eskers, kame terraces, sand plains, and areas of till or ground moraine, which record with great clearness the events of the ice retreat. Many of these deposits have been opened by excavations, and the internal structures thus exposed, if properly interpreted, will give a fairly complete record of how the ice retreated. This record seems to indicate a method of retreat which differs somewhat from any of those thus far formulated. After describing the deposits involved the conditions which seem to have caused them will be outlined.

ESKERS

Typical eskers or serpentine ridges occur in many places and at various elevations in the Nashua Valley. They vary greatly in size and in length, as well as in the coarseness of the component materials. But in every case where the internal structure is well exposed

lished conclusions differing materially from this quotation. In a recent paper he states: "The conception of general deglaciation from north to south rather than from south to north is not supported by the facts. . . . . A third hypothesis must be entertained: that during the deglaciation of Connecticut, the ice margin wasted northward, but that through an unknown distance inward from the periphery, the ice was chiefly stagnant." Richard Flint, "Deglaciation of the Connecticut Valley," *Amer. Jour. Sci.*, Vol. XXIV (1932), pp. 152–56.

[4] S. A. Andersen, "The Waning of the Last Continental Glacier in Denmark as Illustrated by Varved Clay and Eskers," *Jour. Geol.*, Vol. XXXIX (1931), pp. 609–24.

they show a similar composition and arrangement of material. Two individuals are selected for description.

One of these eskers is a long, winding ridge which starts southeast of the village of West Groton and extends southward more than half a mile until it is cut off by the Squannacook River. This ridge rises abruptly 50–60 feet above the level plain on the east. Near the southern bend of the main esker a smaller secondary ridge flanks it on the east, and a kettle or depression filled with water and swamp vegetation occupies the space between the two ridges. On the west

FIG. 1.—Gravel pit in north end of esker near West Groton, showing undisturbed beds in center and slumped material on either side.

an irregular knob-and-basin sand deposit lies in the area between the esker and the river.

The northern end of this esker is used as a gravel pit, and excavation has gone far enough so that the internal structure is well exposed. The material consists largely of gravel with numerous boulders, up to 5 or 6 feet in greatest dimension, scattered through it. The material near the center of the esker is well bedded and undisturbed; that near the sloping margins is disturbed and shows evidence of settling or slumping. At the extreme margin the evidence of bedding is practically destroyed. At the place where this excavation is located the esker rises to a narrow level crest just

over 300 feet in elevation. In the center of the base of the excavation is a ridge or knob of rock, evidently a projecting part of the underlying bed rock. Forty feet of coarse gravel foreset beds, dipping sharply toward the south or in the direction of the axis of the ridge, are superposed upon the rock. Eight feet of very coarse gravel topset beds lie in a horizontal position above the foreset beds. Few of the pebbles in the foreset beds exceed 6 inches in diameter. Many of the larger individuals in the topset beds are 12–15 inches in diameter. Figure 1 shows the end of this esker with undisturbed bedding

FIG. 2.—Picture taken from left margin of pit showing a view at right angles to that shown in Figure 1. This shows bed rock with foreset beds dipping toward the south and horizontal topset beds along the crest of the ridge.

in the center and slumped material at either side. Figure 2, taken from the left-hand margin of the pit, shows a view at right angles to that of Figure 1. This shows the rock ridge (lighter color) at the base, foreset beds dipping toward the left and horizontal topset beds with numerous large pebbles.

The second esker selected for description is located south of South Fitchburg, at an elevation of 500 feet. It starts at the south side of Wanoosnoc Road, follows along parallel to Rodiman Avenue, and for some distance forms the crest of the divide between Monoosnoc Brook and North Nashua River. This esker is also used as a gravel

pit, and a great volume of material has been removed. At one point, near the north end of the pit, bed rock projected up into the esker. The bedding near the middle of the ridge shows little or no disturbance; that near the margins shows settling or slumping with partial

FIG. 3.—Excavation in margin of Rodiman Avenue esker near South Fitchburg, showing ice drag along the margin of the esker.

obliteration of the original structure. On the east side, i.e., on the side of the North Nashua or larger stream valley the gravels also show evidence of slight movement of the adjacent ice after the esker was formed. This movement is recorded by structures similar to that shown in Figure 3, which indicates that it was produced by a

motion of the ice parallel to the direction of the esker ridge. In as far as could be observed in the excavation, this evidence of drag is limited to the eastern flank of the ridge and does not affect the crest or the western flank.

EVIDENCE OF ICE MOVEMENT IN THE NASHUA VALLEY

The evidence of movement, mentioned above, affecting an esker ridge at elevation 500 is the last movement of which any record has been found in the North Nashua River valley. Movement of the ice

FIG. 4.—Bed of boulder-bearing till resting on well-bedded gravels northeast of Lancaster Commons. Arrow indicates line of contact.

mass in the main valley did occur, however, after the 400–420 foot terraces were formed. This movement is recorded by a mass of till deposited when the forward-moving ice overrode the sands and gravels associated with the 400-foot terrace $1\frac{1}{2}$ miles northeast of Lancaster Commons. Figure 4 shows an excavation in which a mass of till varying from 4 to 9 feet in thickness rests upon water-sorted and well-bedded gravels of unknown thickness. The till is typical boulder-bearing material containing many large boulders. Those shown in the foreground of the picture all came from material which has been excavated and removed. The bedded gravels show hardly any disturbance, although there was a little drag along the upper

surface. This is easily seen in Figure 5 where the surface of the gravel is forced up into a little ridge by the overriding of the ice and till.

ORIGIN OF ESKERS

The theory of esker formation advocated in this paper is graphically shown in the three diagrams of Figure 6. The first diagram shows a valley tongue of ice moving forward with a crevasse forming where it passes over a rock ridge near the middle of the valley. The second diagram shows the same crevasse considerably widened with

FIG. 5.—Larger detail of contact between till and gravel showing evidence of drag

water flowing through it and sediment accumulating in the bottom between the walls of ice. The third diagram shows the valley after the last traces of ice disappeared and the crevasse filling is left as a serpentine esker overlying the bed rock ridge. This theory of origin may not explain all eskers, but it will account for those occurring in the Nashua Valley that have been excavated so as to show their internal structure. Bed rock is not present in all eskers, but the internal structure is similar to that described.

MID-VALLEY SAND PLAINS

Another group of deposits, the correct interpretation of which has an important bearing on the conditions prevailing during the waning of the ice sheet, is here discussed under the comprehensive name of

mid-valley sand plains. These sand plains are of three distinct types: (1) long, narrow plains nearly surrounded by ice-contact

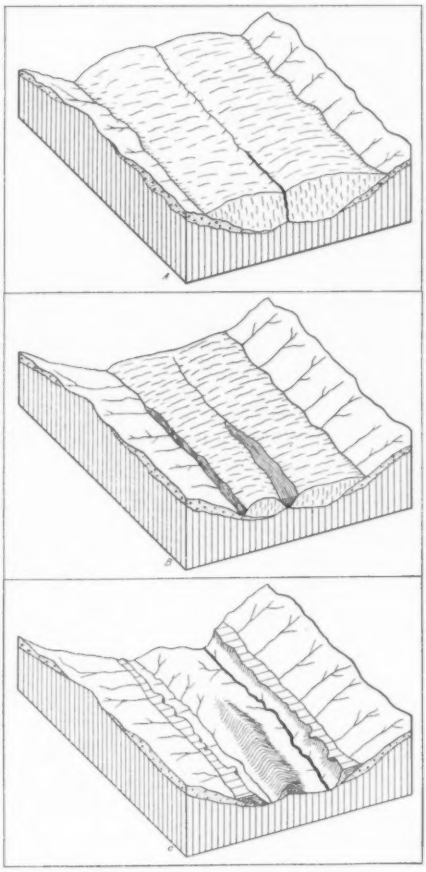

FIG. 6.—Diagrams illustrating theory of esker origin. *A*, valley with tongue of ice in which a crevasse is forming over a ridge of bed rock. *B*, crevasse enlarged and occupied by a stream which is depositing sediment in the bottom of the space. *C*, ice completely melted away and crevasse filling left as an esker ridge.

slopes, with their greatest dimension transverse to the valley; (2) long, narrow plains completely surrounded by ice-contact slopes,

with their greatest dimension parallel to the valley; and (3) broad, level plains, generally surrounded by ice-contact slopes, occupying the central part of the valley with water or lower plains on all sides. Any acceptable theory of ice waning must not only explain the kame knobs and esker ridges within the valleys, and the terraces which fringe their margins, but must also explain these much larger structures which occupy the central areas of the valleys.

TRANSVERSE MID-VALLEY PLAIN

The best example of a transverse mid-valley plain is located in Baker Brook valley, $\frac{1}{4}$ mile north of the Boston and Maine (Fitchburg) railroad. Viewed from the north this looks like the northern margin of an apron plain or gravel type recessional moraine. It is pitted with kettles and fringed with kames, lesser gravel knobs, and ridges. Throughout, its northern margin is characterized by typical ice-contact slopes.

This plain, however, is nowhere more than 200 or 300 yards across. At some places it is much narrower than this. Its southern margin is also an irregular ice-contact slope, with associated kames and lesser gravel ridges, dropping off abruptly to another plain over 40 feet lower in elevation. Moreover, it is continuous across the valley except for a comparatively narrow gap through which the brook now flows, and it is connected on either side of the valley with the kame terraces which fringe the valley for long distances both to the north and to the south. Figures 7 and 8 are typical views of this transverse plain. The former is a view from the north showing the typical ice-contact slopes. In the foreground is a single kame hillock, almost completely detached from the plain proper, yet rising to a level top, with ice-contact slopes on all sides. The latter figure shows a typical view of this plain from the south. The transverse plain is shown in the background, while a lower-level plain appears in the foreground, with a small detached kame knob rising through it near the center of the figure, and a kettle depression, through which the brook flows, lies to the right of this kame. The higher plain is continuous with the kame terrace along the margin of the valley shown at the extreme right.

LONGITUDINAL MID-VALLEY PLAINS

There are two excellent examples of longitudinal mid-valley plains in the Nashua Valley. One is located at the junction of Baker Brook

FIG. 7.—North margin of transverse plain in Baker Brook valley showing ice-contact slopes and a kame in the foreground.

FIG. 8.—South margin of plain across Baker Brook valley with a lower sand plain in the foreground and a kame terrace at the right.

and the North Nashua River near South Fitchburg. The other is near the junction of Squannacook River and the Nashua north of

Ayer. Both are relatively long, narrow plains rising abruptly to a level surface, with either ice-contact slopes or erosion slopes completely surrounding them. Extensive excavations in each of these, made during railroad construction many years ago, show that they are wholly composed of sand and gravel. These excavations are so old that their faces have slumped and it is impossible to see the nature of the bedding.

BROAD VALLEY SAND PLAINS

A number of broad valley sand plains are located at different elevations in various parts of the valley. Two of these northwest of Ayer are selected for description. They are separated from the terraces on either side of the valley, and are completely surrounded by water or by lower land. One of these occupies practically all of the area embraced by the river, Nonacoicus Brook, and the Peterboro branch of the railroad. This area is incorrectly represented on the topographic map as a hill rising over 340 feet, whereas, except for a little granite knob in the southeastern part, it is a broad, level plain approximating 280 feet in elevation. Another smaller sand plain at the same elevation occupies the area on the opposite side of the river immediately south of Squannacook River. Both of these plains have ice-contact slopes on their margins and are surrounded by either water or lower land.

THEORY OF ORIGIN

The esker ridges, narrow valley plains, and broad valley plains seem to represent three distinct steps in the development of structures dependent upon a single cause. The esker ridge represents a deposit formed by a stream flowing in a channel which arose from the widening of a crack or crevasse in the ice over a projecting knob or ridge of bed rock. The continued widening of such a channel accompanied by filling up to a definite water level or water plain would give rise to a long, narrow, level plain such as those described above. Further widening of the space, if accompanied by continued filling with sediment, would expand such long, narrow plains into broad, rounded plains like those found in the valley. This explanation is further suggested by the fact that a prominent granite knob juts through one of these plains near its southern margin.

MID-VALLEY OR RELICT SWAMPS

One of the features of this and many other valleys, which has attracted little attention in the past, is the presence of broad swamp areas along, or near, the channels of the main river and many of its tributaries. Many of these swamps are surrounded or almost surrounded by sand and gravel terraces, yet they certainly do not owe their origin to the erosion of these sediments and later ponding of the river by differential uplift in postglacial time, for they occur in equal abundance on the southward-flowing tributaries and along the northward-flowing master stream. Among the most prominent and characteristic of these swamps are the following: Pigeon Swamp along the North Nashua River between South Fitchburg and Leominster; the great swamp along the Nashua River west of Prospect Hill in Harvard; the large swamp area at the junction of the Squannacook and Nashua in Groton; two swamps along Mulpus Brook, one northwest of North Shirley, and the other east of Lunenburg; and Ash Swamp on the Squannacook near West Townsend. There are many other areas similar to these in every respect, but not large enough to appear on the map.

These swamp areas are partially, or, in some cases, almost completely, surrounded by sand and gravel plains and terraces with ice-contact slopes facing toward the swamps. Many of them show till or boulder clay as underlying material. These areas were never covered with sediments, because the conditions are such that sediments once deposited could not have been removed by erosion. Some of the swamps are filled with peat or other marsh deposits, so that the underlying material is not accessible, but these often show till at, or near, their margins.

The conclusion that these swamp areas were occupied by great blocks of ice while deposition was going on in the valley, and that these ice masses remained in the swamp areas until after deposition ceased, seems to be supported by all of the evidence. When these masses finally did melt away they uncovered the till or ground moraine in the middle of the valley and left the ice-contact slopes around the margin.

THE WANING OF THE ICE

The sequence of events, that would cause the deposits and structures described above, would seem to be as follows. The ice sheet disappeared first over the hills and higher ridges. Long narrow tongues of ice extended along the deeper valleys, and in the Nashua Valley the marginal terraces indicate that at one time a tongue of ice extended from the New Hampshire line to Clinton, a distance of more than 20 miles. In the earlier stages of their existence these tongues of ice were connected to larger masses on the north or on the east. They moved to a limited extent after the side walls of the valley were freed from ice, but eventually they became stagnant and lost all motion. At this stage modified drift was heaped against the valley masses of ice on either side. Occasional crevasses opened across these stagnant masses and became the channels for the drainage water of the valley. This drainage water did not represent great preglacial floods derived from the rapid melting of a continental glacier near at hand, but consisted largely of the normal rainfall runoff from the valley, augmented to a slight extent by the slow melting of the valley masses of ice. As these crevasse channels widened by the gradual melting of the adjacent ice, the increased space was immediately filled by fresh supplies of sediments brought in by the flowing waters. These sediments were also largely derived from the adjacent land and only to a slight extent from the ice mass or from subglacial channels. The land surface, newly emerged from beneath the ice, had little or no protective cover of vegetation and was therefore subject to rapid erosion. Studies of the materials within the deposits show that by far the greater part of it was derived from nearby sources; pebbles of remote origin form but a minute fraction of the total. In some parts of the valley this process of widening by melting and filling with sediment continued, for a long time, until the broad mid-valley sand plains were formed.

When at length the melting of the ice exceeded the filling by transported material, more spacious channels were opened and lower outlets uncovered. This allowed the ponded waters of the valley to drain away, and, as a result, the remaining masses of ice melted at an increased rate. The melting out of these final masses from the

deeper parts of the valley gave rise to the relict swamps which are characterized by a poorly drained or undrained surface and the absence of water-sorted sediments.

Andersen[5] has suggested that esker ridges, kames, and pitted sand plains were all formed under the ice and not upon its upper surface or between masses of ice. He states that:

> The topographic expression of pitted plains and kames with undisturbed layers does not originate from a collapse of the whole deposit, but the surface is largely a cast of the under surface of the ice at the time when the deposition ceased. Accordingly the ice in the kettles must have been covered by a rather thin veneer of drift, and the ice-contact slopes must be regarded as due to a slumping of gravel only in cases where such disturbance can actually be proved. In most cases it appears that the present slopes are largely the original ones.

In central Massachusetts all of the evidence is opposed to the theory advocated by Andersen. Slumping of the ice-contact slopes is present along the margins of the eskers, around the kame hillocks, and along the valley margins of kame terraces.[6] Some of the smaller kames which do not rise to the full height of the associated terraces show by their internal structure that they were formed at the surface of the ice and let down, when it melted, upon the underlying surface. This is clearly shown by the presence in the central parts of these of minute faults, some of which show a movement of a few inches, and others only a fraction of an inch. This disturbance of the interior is associated with much more extensive slumping of the ice-contact marginal slopes.

CONCLUSIONS

In the central Massachusetts region the ice sheet when waning disappeared first from the higher areas and left long narrow tongues in the deeper valleys. These tongues of ice were affected by limited movements after many of the areas of modified drift were deposited. Eventually, however, these valley masses became severed from the main ice sheet and were then dissected by enlarged crevasses. Water, largely derived from the natural drainage of the valley, and carrying

[5] *Op. cit.* (1931), p. 613.

[6] T. C. Brown, "Evidence of Stagnation during Deglaciation of the Nashua Valley," *Amer. Jour. Sci.*, Vol. XIX (1930), pp. 359–67; "Kames and Kame Terraces of Central Massachusetts," *Bull. Geol. Soc. Amer.*, Vol. XLII (1931), pp. 467–79.

with it copious sediments also largely derived from the surrounding land, flowed into and through these crevasses, forming within them the esker ridges, valley plains, and other stratified deposits which now characterize the valley. The final masses of ice, when they melted out, left till-covered areas in the deeper parts of the valley which are now swamps. The stagnant masses of valley ice at their maximum extent stretched from the vicinity of Clinton to the New Hampshire line, thus indicating a broad stagnant border zone.

GENERAL GEOLOGY OF THE CRYSTALLINES OF ALABAMA[1]

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ABSTRACT

A reconnaissance mapping of the crystalline area of Alabama has been completed. The so-called Archean of earlier reports has been divided into belts. Some Triassic diabase dikes in Alabama have been mapped. A tentative interpretation of the structure of the crystallines and the relations of the more important intrusives are here presented. The field work has been correlated with mapping in progress in Georgia and is offered as a basis for more detailed studies.

INTRODUCTION

The crystalline rocks in Alabama are of special interest because they are the southwesternmost portion of this element of the Appalachian region. From deep well records it is known that they extend for a considerable distance to the southwest under the Upper Cretaceous formations. It has been maintained that because the Appalachian and Ouachita Mountains are similar in structure and age they are continuous under the blanket of formations which occupy the Mississippi embayment,[2] but it should be noted that the crystallines of Alabama are not represented in the Ouachita Mountains, at least for the most part. The belt of folded Paleozoics in Alabama is narrowed to about 30 miles, while the width of the crystallines across the strike is about 85 miles. Without evidence as to the westward extension of the crystallines there is much uncertainty as to the relations of the Appalachian and Ouachita structures.

RECENT FIELD WORK[3]

The general geology of the crystallines, excepting the Talladega series, which was described by Dr. Charles Butts, was discussed by

[1] Published with the permission of the Geological Survey of Alabama.

[2] For a recent discussion of this problem see W. A. J. M. Van Waterschoot van der Gracht, "Permo-Carboniferous Orogeny in South-Central United States," *Bull. Amer. Pet. Geol.*, Vol. XV, No. 9 (1931), pp. 991–1057.

[3] The interpretation of the crystallines which is here set forth is based on the writer's work and the opinions are his own, but he desires to acknowledge his indebtedness to Dr. Jonas and Dr. Crickmay for helpful criticisms and suggestions, and to Dr. Charles Butts for the description of the Talladega series, which the writer has reviewed in the field. It is probable that the interpretation here presented will be modified when further work is done and the difficult history of the crystallines is studied in detail.

the writer in 1926 from meager information, and the location of the rocks was shown in a general way on the geologic map of the state published that year,[4] with due acknowledgments of the sources of the information. In the report on the gold deposits of Alabama[5] the writer published a sketch map which shows some corrections and additions made during the progress of field work in 1929.

While working in the summer of 1930 for a report on the mica deposits, the crystallines were reviewed in part, and it was found that the area shown as "Archean" on the state map in 1926 was susceptible of division and that a better interpretation of the area was possible. Field work was accordingly continued in the summer of 1931, and some discoveries were made which add to our knowledge of the geologic history of the area. The newly recognized divisions of the crystallines were mapped, and a more detailed discussion of the geology is now offered.

The Geological Survey of Alabama secured the co-operation of Dr. Anna I. Jonas, who is working on the crystallines of the Appalachian region for the U.S. Geological Survey, and a brief inspection tour of the crystallines of Alabama was arranged for her in December, 1930.

During the summer of 1931, with the sanction of the state geologists of Georgia and Alabama, Dr. G. W. Crickmay, of the Georgia Survey, and the writer co-operated in the study of the rocks along the Alabama-Georgia state line.

GENERAL DESCRIPTION

The crystalline rocks of Alabama which occur in belts or zones extending in a general northeastward and southwestward direction are a part of the Piedmont Upland which extends to the northeast through Georgia and are an integral part of the Appalachian region which terminates southward, in so far as surface features are concerned, in eastern Alabama. Their northwestern border is overthrust to the northwest on unaltered Paleozoics. To the southwest they are covered by the Upper Cretaceous formations.

[4] See *Special Report 14* and state map prepared in co-operation with the U.S. Geol. Surv., Geol. Surv. Ala. (1926).

[5] "Gold Deposits of Alabama," *Geol. Surv. Ala. Bull. 40* (1930).

The divisions of the crystallines which are mapped (see Fig. 2) are based on structure and the kinds of rocks. They serve particularly as a basis for the interpretation of the general succession of geologic events. However, the description of the divisions proceeds

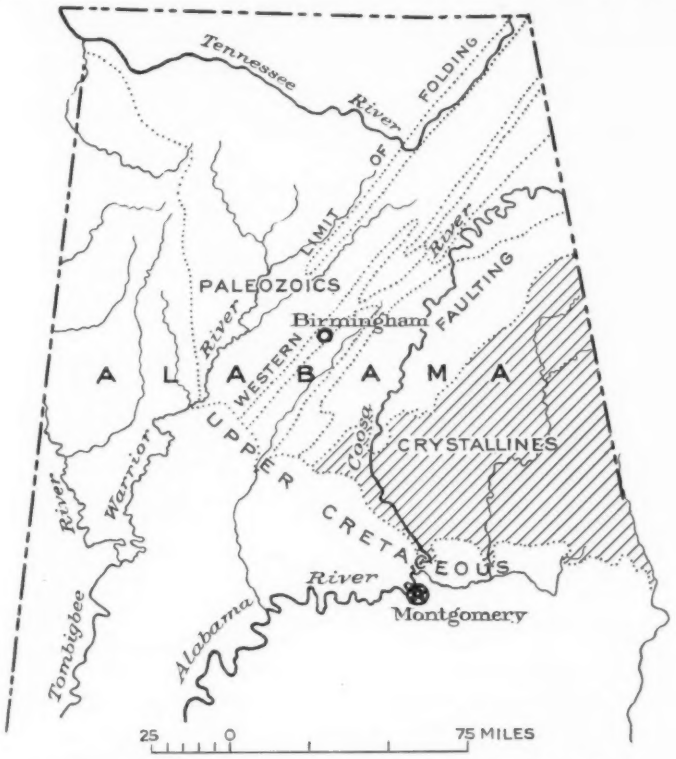

FIG. 1.—Sketch map showing the relations of the crystallines in Alabama

from the northwest to the southeast, following in general a geographic rather than a geologic order.

TALLADEGA SERIES

This series of more or less metamorphosed sediments, which possibly aggregates 30,000 feet, contains slates and sericitic phyllites in

which are interbedded, at certain localities and horizons, conglomerates, sandstone, limestones, dolomite, marble, quartzites, chert, graphitic phyllite, and quartz schists. Some of these have been mapped and described as members of the series by Butts.

The northwestern border of the Talladega area, in most places, presents a rather bold escarpment, which rises above the general

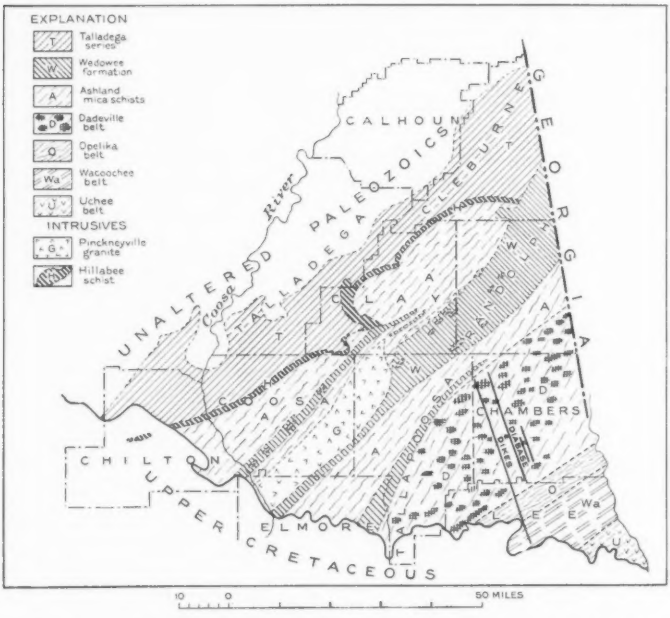

FIG. 2.—Sketch map showing the divisions of the crystallines in Alabama

surface of the unaltered Paleozoics. The line of separation between the Talladega and the unaltered Paleozoics is the great thrust plane on which the former has moved as much as 15 miles northwestward. This estimate is based on the positions of the formations with respect to the fault plane in the southern portion of the field where window-like (fenester) areas of the unaltered Paleozoics occur. Formerly the Talladega must have extended much farther to the northwest, but it has been removed by erosion.

The belt is very irregular in form and varies in width from about 8 to 22 miles. The cleavage dips in the Talladega series are mostly 40° or more to the southeast. There is evidence of repetition by folding and thrust faulting. Rocks which are present in the southern part of the belt and are conspicuous members of the series there, are absent or not exposed in the northern part. Certain phases in the northern part were regarded by Dr. Jonas as phyllonitic.

TABLE I

TALLADEGA SERIES

CHILTON AND COOSA COUNTIES		CLAY COUNTY	
Thickness in Feet	Members	Thickness in Feet	Members
		1,000	Phyllites, etc.
		2,000	Erin shale (Pennsylvanian?)
		2,000	Phyllites, etc.
		20	Dempsey marble
		1,000	Phyllites, etc.
500	Yellow leaf quartz schist, Jemison chert } Lower Devonian		
1,000	Butting Ram sandstone	?	Cheaha sandstone
7,500	Phyllites, etc.		
100	Jumbo dolomite		
4,500	Phyllites, etc.		
200	Quartzite		
3,000	Phyllites, etc.		
100	Ferruginous sandstone		
1,000	Phyllites, etc.		
1,000	Brewer phyllite		
500	Phyllite, etc.		
100	Sawyer limestone	?	Sylacauga marble
4,500	Phyllites, etc.		

The southeastern border of the Talladega is marked for nearly 100 hundred miles by the Hillabee schist, which is an intrusive sheet or sill occurring along a great overthrust plane. In the northern part of the field there is a fault which brings the Wedowee in contact with the Talladega.

The age of the Talladega ranges from probable pre-Cambrian to Pennsylvanian. It is not discussed in detail here, but the following general section, compiled from the description by Butts, will give an idea of its character and the relation of its members in the southern part of the belt. In the northern part of the belt no division

into members has been made and few of those in the southern part can be recognized.

No igneous rocks have been recognized in the Talladega, though it contains some auriferous quartz veins which have been worked.[6]

HILLABEE CHLORITE SCHIST

This chlorite schist is interpreted as a sill of altered igneous rock which was intruded along, and apparently lubricated, a great thrust plane. It extends with some short interruptions for about 100 miles and lies between the Talladega series and the Ashland mica schist which has overridden the Talladega from the southeast. In its most expanded area it is probably more than 500 feet thick. Coarse-grained specimens from it show the structure and minerals of a diorite. There are two small inliers of the Hillabee in the Upper Cretaceous (Tuscaloosa formation) and it possibly continues a considerable distance farther to the southeast under cover of these Mesozoic sediments.

In places the Hillabee is attenuated or absent for short distances. At its northeastern extremity it is cut off by a thrust fault beyond which it has not been recognized. Auriferous veins and pyrite deposits in the Hillabee have been worked.[7] The metallization was subsequent to the development of the schistose structure, since the veins cut across the planes of schistosity and have the characteristics of stringer veins. In the Ashland mica schists there are intrusions which, adjacent to the northern part of the Hillabee, show a gradual transition to non-schistose diorite. On the map of Clay County, Prouty has shown an area of these rocks,[8] and the writer has noted them at other places along the northwestern border of the northern area of the Ashland, but they have not been mapped.

The relations of the Hillabee to the Carboniferous (Erin shale) portion of the Talladega imply post-Carboniferous age for the intrusion of the Hillabee and the related diorites.

[6] "Gold Deposits of Alabama," *op. cit.*, pls. 2 and 4.

[7] *Ibid.*, pls. 1, 2, and 3.

[8] W. F. Prouty, "Geology and Mineral Resources of Clay County," *Geol. Surv. Ala.* (1923).

ASHLAND MICA SCHISTS

These schists are metamorphosed sediments apparently of the pre-Cambrian age.[9] They contain no distinctively calcareous beds and are dominantly of two types: garnetiferous biotite schists and muscovite schists. There are also variable bands of graphitic schists, in which the graphite occurs as crystalline flakes. The geographic distribution of the graphitic schists, which are found principally in the western part of the area of the Ashland, is apparently due to lateral variations in the carbonaceous content of the sediments from which they were derived. Siliceous schists, quartz schists, and micaceous quartzites occur in the Ashland. They resist erosion and in the western belt of the Ashland have given rise to ridges which, by their trends, emphasize the structure of the Ashland and imply repetition by folding and extensive overthrusting to the northwest. The thickness of the Ashland has not been determined, but the opinion has been ventured that it is much more than 10,000 feet.

The typical Ashland has been intruded locally by dioritic rocks along the northwestern border near the Hillabee sill, as has already been noted in discussing the Hillabee. It has also been intruded by the Pinckneyville granite, to be described later. In the Ashland there are many bodies of pegmatite which the writer believes are related to the regional magma which produced the Pinckneyville granite, and it is in these pegmatites that most of the mica deposits of Alabama are found.

The metamorphism of the Ashland has been locally of such nature as to give rise to kyanite schists and masses of kyanite, which, however, have not yet been found in workable deposits. Garnet zones occur locally. Some tourmaline occurs, particularly in the quartz veins. Staurolite schists and knotted schists are also present.

Besides the mica deposits which occur in the northwestern and southwestern areas of the Ashland crystalline graphite has been mined in the belts of graphitic schists mentioned above, and auriferous quartz veins have been worked at many places.[10] The gold deposits

[9] These rocks were classified as Algonkian in *Special Report 14, Geol. Surv. Ala.*, p. 32, but the writer does not wish to follow the terminology, Algonkian and Archean, as was used by the U.S. Geol. Surv., because of the uncertainty of correlation.

[10] See *Bull. 40, Geol. Surv. Ala.*, pls. 1, 2, and 4.

have been explained by the writer as having been formed subsequent to the intrusion of the Pinckneyville granite. The metallization was probably from the general regional magma which gave rise to quartz veins at the close of the Paleozoic after the Appalachian structure was developed.

In the western portion of the southeastern belt of the Ashland adjacent to the Wedowee there are some rather conspicuous granite gneisses which are perhaps ancient granite intrusions that have suffered more than one period of deformation. They are unlike any other rocks here described and may be related to gneisses at other places in the formation which weather to billet-shaped masses.

Particularly there should be mentioned the occurrence of hornblende augen-gneiss along the southwestern border of this area of the Ashland. On the state map the area in which this gneiss is found was distinguished by a special pattern. It occurs in irregular intrusions and in some places weathers to large boulders. Usually, however, it is found only in small exposures. Locally, the pink orthoclase feldspars in this gneiss are as much as 2 inches in length. When the gneiss weathers, the partially decomposed feldspars, which may be picked out of the resulting clay, frequently show the form of Carlsbad twins. Usually the feldspars are small, the augen-gneiss has suffered pressure, and the feldspars are sheared into elongate "eyes." The age relations of this gneiss are not clear. It has not been seen to cut the Wedowee formation which borders it on the southeast, but it may be younger than the Wedowee.

WEDOWEE FORMATION

This formation is phyllitic, but contains slates and quartzites. No distinctively calcareous beds are seen in the typical Wedowee in Alabama. The formation is characterized in many places by the occurrence of amorphous graphite, which renders the rocks black or grayish black. The graphite also occurs as impure friable beds.

The rocks which constitute the Wedowee were presumably deposited as sediments of different sorts, a part of which were carbonaceous and locally highly charged with such material. In the southeastern belt of the Wedowee the lower part of the formation is characterized by quartzites, which are well exposed in the railway

cut at Jackson's Gap and the outcrop in the southern part of the ridge known as the Devil's Backbone.

The Wedowee was apparently deposited on the Ashland, since it is infolded or by thrust faulting is imbricated with these schists. The large area of the formation on which the town of Wedowee is located contains belts of schists which have the characteristics of the Ashland schists, but it has not been possible to do sufficient field work to differentiate them in the mapping.

The Wedowee does not contain the dioritic intrusives which are characteristic of the Ashland, or at least these intrusives are generally absent. There are amphibolites in the southeastern part of the main area of the Wedowee, but these are believed to occur in the undifferentiated belts of Ashland which are included in the Wedowee in the general mapping.

Within the Wedowee there are intrusions of the Pinckneyville granite, some of which are exposed as "flat rocks" of considerable extent. There are also minor intrusions of the granite which have risen parallel with the schistosity of the formation. These are too small to be shown in the general mapping. The occurrences of granite show a southwest-northeast alinement which is related to the structure of enclosing rocks.

The thickness of the Wedowee is difficult to determine. It may comprise more than 10,000 feet of metamorphosed sediments. The black graphitic phase reaches more than 2,000 feet in places, if there is no repetition by faulting where the observations were made.

The age of the Wedowee has not been definitely determined. A portion of it has been tentatively assigned to the Carboniferous. It may range into the Cambrian, since the lower limit of the formation is indefinite. The Carboniferous age of the upper part is supported by its similarity with the Carboniferous portion of the Talladega, the age of which has been determined by fossils.[11] The Wedowee, together with the Ashland, may be the equivalent of the Talladega.

PINCKNEYVILLE GRANITE

The principal area of this granite lies in the southwestern part of the field and is from 6 to 8 miles wide and about 40 miles long.

[11] *Special Report 14, Ala. Geol. Surv.* (1926), pp. 217–19.

To the northeast the granite is found intruded between the Wedowee and Ashland and in the Wedowee along a fault zone. There are also bosses, or "flat rock" areas, in the Wedowee, some of which are shown on the map. In the northwestern area of the Ashland the granite occurs as minor intrusions, mostly as thin sheets or stringers in the schists. The small intrusives and the borders of the larger masses are gneissic. The age of the Pinckneyville granite has generally been considered post-Carboniferous. From its relation to the structure and to the Wedowee in which it is intrusive, it is the youngest granite in the crystallines of Alabama. Pegmatites which are apparently related to this granite are conspicuous in the area of the Ashland and are the source of the mica which has been mined in the western districts of the crystallines.

DADEVILLE BELT

This term is here introduced to designate a division of the area shown on the state geological map as principally Archean.[12]

The oldest portion of the Dadeville belt or complex consists of rocks which are in general like the Ashland, but they are largely obscured by intrusives. The older intrusives include basic rocks with conspicuous hornblende which commonly weather to form what is termed in the region "brick bat," and the soil where they occur is dark red. Younger than these basic rocks and cutting them are granitic intrusives. The distribution of the larger and more conspicuous granitic areas is shown in a general way on the sketch map. The granite is here named the Sougahatchee granite from the creek near which it is exposed at many places in the southwestern border of the area of the Dadeville belt. The basics which have not been differentiated on the general map include hornblende diorites, amphibolites, hornblende olivine rocks (Cortlandtite), diabases, gabbros, and pyroxenites.

The Sougahatchee granite forms "flat rocks" at many places, and where not well exposed its presence may be inferred from the sandy, light-colored soil which is in decided contrast with that formed by the weathering of the schists and basics. The Sougahatchee granite

[12] Since the terms "Archean" and "Algonkian" cannot now be applied satisfactorily in Alabama, the rocks which are older than the Paleozoic are here called pre-Cambrian.

differs from the Pinckneyville granite in being much more gneissic. It contains microcline while the Pinckneyville granite does not, but shows oligoclase.

In the Dadeville belt north of Dadeville, there is a zone passing near the old post-office of Easton, in which corundum and asbestos were prospected some years ago. In this zone the writer discovered chromite in the summer of 1930. It may be found on the surface, having weathered out of the containing rocks, but its genetic relations have not been determined, since there is deep weathering, and no prospecting for it has been done.

The Dadeville belt has not been found to carry auriferous veins, and they are also absent from the belts of the crystallines farther to the southeast in Alabama. The Dadeville mica district lies near the western border of this area in a portion of the complex where schists and pegmatites are conspicuous.

OPELIKA BELT

This belt is about 5 or 6 miles wide, extending in a northeast-southwest direction from the vicinity of Opelika. In it there are conspicuously exposed schists which are similar to the muscovite schist phase of the Ashland. It also contains belts of biotite gneiss. There are a few occurrences of granite gneiss. This area is differentiated from the Dadeville belt by the absence of the basics and the presence of the biotite gneiss.

Within a zone of biotite gneiss which lies about 3 miles to the northwest of Opelika, there are some muscovite mica prospects.

WACOOCHEE BELT

Southeast of the Opelika belt is an area containing schists, two quartzite zones (Hollis quartzites), two well-defined zones and some minor occurrences of banded porphyroclastic rocks (mylonites), a zone of granite gneiss, and lenses of biotite augen-gneiss.[13] Adjacent to the southwestern ends of the two conspicuous quartzite zones are some small outcrops of dolomite (Chewacla dolomite). The interpretation of these rocks, which have been closely folded and subjected to intense thrusting, is manifestly difficult. The schists

[13] Dr. Jonas regarded this gneiss and associated fine-grained aphanitic layers as mylonitized granite.

may be principally the non-muscovitic phase of the Ashland, but they carry, at certain places, amorphous graphite and, in one place, a narrow bed of this material which is suggestive of the Wedowee.

The quartzites, which have been called the Hollis quartzites,[14] are dense to micaceous and in places grade into beds which, as exposed, have the appearance of sandy and clayey shales. In other localities they are interbedded with somewhat muscovitic schists. These quartzites have a conspicuous development in Georgia in Pine Mountain and Oak Mountain. They have been discussed by the writer,[15] who considered them and the associated dolomites as possibly of Paleozoic age and suggested a correlation with similar Paleozoic formations in the Talladega described by Butts at localities some 60 miles to the northwest, but it is impossible to make any certain correlations without fossils or physical continuity. The quartzites may be the equivalent of quartzites in the Ashland and, if so, would be of pre-Cambrian age according to the usage in this report.

The dolomite, which is known as the Chewacla dolomite, is most extensively exposed in the old quarry at the locality from which it is named, but its relation to the Hollis quartzite cannot be seen there and has not been observed at any other place. At four localities where it is found the dolomite lies to the southeast of the quartzite and presumably below it, but its stratigraphic position is not certain. One locality of the quartzite is an inlier in the Upper Cretaceous, and there the dolomite is not exposed, but its presence is suggested by the occurrence of large springs near-by. It is peculiar that the dolomite exposures are all close to the southwestern border of the crystallines—a fact which suggests their deposition in the arm of a sea which was open to the southwest—while the extensive development of the quartzites is to the northeast.

The best exposure of the banded porphyroclastic rocks (mylonites) is found in the southeastern belt at Bleecker. The writer collected a specimen at Bleecker in 1925, but the rock exposed at that time was non-porphyritic and so dense that its character was not determinable in the thin section. Since then an excavation for the end of a railway "Y" has cut through the rocks, and they may be

[14] *Special Report 14, Geol. Surv. Ala.* (1926), p. 33.

[15] "Significance of the Quartzites of Pine Mountain in the Crystallines of West Central Georgia," *Jour. Geol.*, Vol. XXXVIII (1930), p. 271–79.

seen to good advantage. The banded phase of these rocks in Alabama was found by the writer in the summer of 1931, at a point about ½ mile northwest of Wacoochee.[16] The rocks vary from dense beds to thin, banded porphyritic beds. At localities in the northeastern part of the Wacoochee area they appear to grade into biotite augen-gneiss, the finer-grained phase of which resembles the banded porphyry (mylonite).

The most interesting exposure of the biotite augen-gneiss is found at the bridge over Chewacla Creek about 4 miles due south of Auburn. The north end of the bridge is on the augen-gneiss, and the south end on the Hollis quartzite. Large pieces of the augen-gneiss, which have been blasted away in making the road to the bridge, show porphyroclasts of feldspar which are as much as 2 inches in diameter. Many of the phenocrysts are rounded in section; others are typical "augen." In the bed of the stream the gneiss may be seen intruded between the beds of quartzite. At other localities the gneiss is of finer grain and is found in the schists, or near the quartzites.

The exposures of granite gneiss (mylonite) are found principally between the two well-defined belts of banded porphyries. The important outcrops which have been seen in this zone are of a tough dark rock, which, however, weathers to yellowish boulders. Near Wacoochee there are some small outcrops of a grayish granite gneiss and on the Chattahoochee, near Goat Rock dam, granite mylonite occurs.

UCHEE AREA

The most southeastern division of the crystallines contains some good exposures on Little Uchee Creek and its tributaries from the North. There are also good exposures in the bed of the Chattahoochee River below the dam between Phenix City, Alabama, and Columbus, Georgia. The geology of this area is largely obscured by remnants of the Upper Cretaceous (Tuscaloosa formation) and the river terrace gravels. The northwestern part of the area contains granite gneiss. On Uchee Creek at Moffet's Mill there is granite gneiss and biotite gneiss. On White's Creek a mile and a quarter to the east there is a good exposure of contorted biotite gneiss in-

[16] The interpretation of these rocks as mylonites was suggested by Dr. Crickmay, who has since found localities in Georgia where there is a gradation from the mylonites, to the augen-gneiss and has confirmed this conclusion by petrographic studies.

jected by granite, which may be seen from the highway bridge. There is some hornblende gneiss along the Opelika-Columbus highway; and in the bed of the river between Phenix City and Columbus, contorted hornblende gneiss injected by granite is seen to advantage. Whether or not schists occur in this belt was not learned from the exposures which were seen.

DIABASE DIKES

In the summer of 1931 the writer discovered a diabase dike $2\frac{1}{2}$ miles northeast of Auburn. Later he traced it and found that it extends about 33 miles from the border of the Upper Cretaceous in Lee County to the valley of the Tallapoosa River in the northwestern corner of Chambers County. The strike varies from about 8° to 15° west of north. This dike is here named the Auburn dike. Boulders from it may be found about 2 miles east of Auburn on the highway to Opelika. The dike averages about 50 feet wide, but narrows in places to perhaps 20 or 30 feet. It is thought to be continuous, since nineteen outcrops were found, well spaced along its extension. No doubt more outcrops could have been found if time had warranted the search. One of the good exposures which may be readily visited is situated about $\frac{3}{4}$ mile east of Stonewall, a station on the Central of Georgia Railway. Conspicuous boulders from this dike, which have long attracted attention because of their toughness, may be seen near the residence of D. S. Robertson, Jr., about $2\frac{1}{2}$ miles west and a little north of Oak Bowery, Cherokee County, in the southeast quarter of Sec. 31, T. 21 N., R. 26 E.[17]

A little later in the season another dike was found just east of the old postoffice site of Marcott in Chambers County, which is about 6 miles northwest of LaFayette. This dike, which is here called the Marcott dike, is parallel with the Auburn dike and lies about $3\frac{1}{4}$

[17] In 1838, Toumey stated on page 65 in his *Report on the Geology of South Carolina*, that he had traced the basaltic or trap dikes of Virginia and North Carolina through Georgia and as far as the Coosa River in Alabama. This has been quoted as evidence of Triassic dikes in the region. Toumey in his subsequent reports on Alabama did not describe any such dikes. Evidently he did not see any diabase dikes in Alabama.

It is probable that it was from the Auburn dike that Dr. Eugene A. Smith, former state geologist, collected a specimen about 3 miles southeast of Auburn. See *Bull. 5, Geol. Surv. Ala.*, Part 2 (1869), pp. 145–46. J. Morgan Clements sought for the dike but did not find it. He, however, described the specimen collected by Dr. Smith as a typical diabase. Dr. Smith appended a note to the effect that it was this same rock which Dr. George W. Hawes identified as dolerite in 1875.

miles east of it. It was traced northward to the lowland on High Pine Creek within 2 miles of the Tallapoosa River and southward to an exposure in the southeast quarter of Sec. 2, T. 20 N., R. 26 E., which lies 4 miles north and a little west of Opelika. If the dike continues southward from this exposure it passes through the western part of Opelika. Its known extent is about 26 miles. Twelve exposures of it were found within this distance, and more could be easily located. The thickness of this dike is not so great as that of the Auburn dike. Probably it averages less than 20 feet.

A third dike, here called the Snapper dike from a small creek which it crosses, is known to extend for 3 miles and lies a little less than 2 miles to the east of the Marcott dike. A fourth dike which may be called the Danway dike is exposed just east of Danway siding (formerly Tuckersburg) on the Roanoke branch of the Central of Georgia Railway between Opelika and LaFayette. It is probably at least 6 miles long and lies a little more than $\frac{1}{2}$ mile east of the Snapper dike. The exposure just east of Danway is good and shows the dike to be a little more than 8 feet wide. The Snapper dike and the Danway dike, both of which lie near Danway, could not be easily traced because of the creeks and cultivated fields. Since they are relatively narrow they may not extend far. There may be other dikes, but no special search was made for them, the time for field work being limited.

There are certain diabase rocks described from Alabama which conform with the Appalachian structure; but the dikes, described above, cut across the strike of the crystallines, are later than the Appalachian structure, and older than the Upper Cretaceous. They are presumably contemporaneous with the trap dikes of Connecticut, the Palisades of the Hudson, and other occurrences from Nova Scotia as far south as North Carolina, in which diabase intrusions are found associated with the Newark series. In the southern border of North Carolina, the northern border of South Carolina,[18] in Georgia,[19] and in Alabama, the diabase dikes are not associated with Triassic sediments. Whether or not Triassic sediments were once present in these states is an interesting speculation.

[18] A. G. Keith, "Kings Mountain," *U.S. Geol. Surv. Folio No. 222* (1931).

[19] S. W. McCallie, "A Preliminary Report on the Road Material of Georgia," *Geol. Surv. Ga. Bull. No. 8* (1901).

FISH FAUNA OF THE MIDDLE PHOSPHORIA FORMATION

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ABSTRACT

An unusual group of fish remains occurs in the middle part of the Phosphoria in a Permian member. The fauna presents some notable contrasts with the previously described Pennsylvanian fish fauna of a lower member of the formation. Two new genera and three new species of sharks are described.

The Phosphoria formation is a thin formation of Pennsylvanian and Permian age which occurs in western Wyoming and in adjacent parts of Colorado, Utah, Idaho, and Montana. In the Wind River and Owl Creek Mountains of Wyoming, the area of this investigation, the Phosphoria is underlain by Tensleep sandstone and is overlain by the Dinwoody formation.

Fish remains were discovered in the Phosphoria formation by the St. John expedition and reported in 1883.[1] The fauna was described by E. B. Branson in 1916.[2] All the fish specimens of these reports were collected from a bed of rock phosphate about 50 feet from the base of the formation, the Lower Phosphate member of the writer.[3] Collections of the writer in connection with his work on the fauna of the Phosphoria formation, and especially since its publication,[3] have added a considerable fish fauna from a thin phosphate bed at the top of the Pustula member. This member is in the middle part of the formation and is considered to be Permian in age, as it contains *Spiriferina pulchra* and *Aulosteges*. The Pennsylvanian age of the Lower Phosphate member was established by Branson in 1916.[2]

Of the eleven species of fish recognized in the Pustula member, six are also found in the Lower Phosphate member, and there are ten

[1] O. St. John, "Report on the Geology of the Wind River District," *U.S. Geol. and Geog. Survey Terr., Ann Rept. 12*, Part I (1883), pp. 173–270.

[2] E. B. Branson, "The Lower Embar of Wyoming and Its Fauna," *Jour. Geol.*, Vol. XXIV (1916), pp. 639–64.

[3] C. C. Branson, "Paleontology and Stratigraphy of the Phosphoria Formation," *Univ. Mo. Studies*, Vol. V, No. 2 (1930).

species of the lower member not found in the upper. The species common to the two members are *Helodus subpolitus* Branson, *Cladodus occidentalis* Leidy, *Janassa unguicula* Eastman, ? *Campodus variabilis* Newberry and Worthen, *Ctenacanthus browni* Branson, and spine denticles. Of these species, only two are of value in age definitions, as *Helodus subpolitus* and *Ctenacanthus browni* are found only in the Phosphoria formation and *Campodus variabilis* is doubtfully identified in the lower member. *Cladodus occidentalis* occurs in the Upper Coal Measures of the Mississippi Valley and in the Permo-Carboniferous of Nebraska. *Janassa unguicula* occurs in the Upper Pennsylvanian of Illinois, and the genus is abundantly represented in the Permian.

The following species are confined to the Lower Phosphate member: *Deltodus mercurii* Newberry, *Helodus rugosus* Newberry and Worthen, *Crassidonta stuckenbergi* Branson, cochliodont undetermined, *Janassa unguiformis* St. John and Worthen, *J. angularis* Branson, *Ctenacanthus amblyxiphias* Cope, *C. obscuricostatus* Branson, *Eunemacanthus keytei* Branson, *Batacanthus gigas* Branson. Especially significant is the lack of *Deltodus mercurii* and *Crassidonta stuckenbergi* in the upper bed. These forms are very abundant in the lower bed and are the chief basis for correlation of that bed with the Pennsylvanian.

The species confined to the Pustula member are *Dolophonodus uncinatus* C. Branson, *Hamatus phosphoriensis* C. Branson, *Ancistriodus multisectus* C. Branson, *Edestus* ? sp., and *Ctenacanthus mutabilis* C. Branson. None of these species is found elsewhere. The introduction of such radically different forms as these indicates a considerable stratigraphic break, and supports the writer's contention that the Pennsylvanian-Permian boundary lies between the two fish-bearing members of the Phosphoria formation.

SYSTEMATIC PALEONTOLOGY

Dolophonodus C. Branson, n. gen.

Teeth high-crowned, somewhat recurved, laterally compressed, tip triangular, compressed in anterior-posterior direction, set off from lower part of crown by distinct shoulders, lateral face of crown with oblique row of blunt barbs.

Dolophonodus uncinatus C. BRANSON, N. SP.

Tooth laterally compressed, crown high, pointed, recurved about 20°, terminated by a triangular flattened tip emarginated below by distinct shoulders; anterior corners of tip base with pair of blunt barbs which originate one-third of the way up the tip and diverge so that the free portion of each barb is near the lateral angle of the tip. Each barb followed below by a row of seven similar barbs, the next pair at the lateral margins of the anterior face of the crown below the arrow-like tip, succeeding pairs progressively farther posterior on the lateral faces of the crown, the lower pair at the lateral margins of the posterior face of the crown and supplemented by at least twelve smaller barbs on the base of the posterior face, irregular in size, shape, and arrangement. Each of the second pair of barbs is upon a distinct ridge which at the fourth pair of barbs is separated from a stronger ridge which emarginates the anterior face of the crown by a sharp groove which widens to the base of the tooth. The crown is compressed laterally below the tip, which has sharp lateral edges.

The surface of the tooth is smooth except where fine longitudinal striations may be discerned. At the base of the anterior face of the crown is a triangular roughened area marked by 16 ridges, those in the center of the area 5 mm. long, lateral ridges regularly shorter to the extreme pair which is at the anterior edge of the lateral faces. The ridges are irregular in shape and are separated by broad shallow grooves.

The base of the tooth is longer than the crown, and appears to extend some distance anteriorly—an adaptation which would react against the leverage of an object caught on the tip of the tooth. The dentine core in the base is pear shaped, the larger part anterior. The core extends nearly to the tip of the crown. The ridges at the base of the anterior face are preserved only in the paratype, which also shows indistinct serrations on the lower part of the edges of the tip. The paratype differs from the holotype in that the barbs of the second and third pairs are not opposite each other but alternate on the two sides of the crown. The holotype is 28.5 mm. high with part of the base missing, height of tip 7 mm., width of tip at shoulder 3 mm., width of base of crown 3.5 mm., length of base 5 mm. The paratype

is 27.2 mm. high with part of the tip and base missing, width of tip at shoulder 4 mm., width of base of crown 4.3 mm., length of base 5.6 mm.

No other shark teeth are barbed, although the marginal serrations of some of the Selachian family Charchariidae are somewhat like barbs in that they are at right angles to the direction of the bite. The Phosphoria genus must have preyed upon smaller fish, which it literally harpooned. The position of the teeth in the mouth is unknown, but may have approached the edestid condition of a median symphysial row. The grasping nature of the teeth probably was a disadvantage in that teeth may have been pulled out at too rapid a rate to be replaced effectively.

Locality—Dinwoody Creek, Wind River Mountains, Wyoming.

Hamatus phosphoriensis C. BRANSON

This genus and species has been described in a previous paper. Subsequent collections have shown that it occurs in abundance in the Pustula member. Thin sections reveal a lateral fibrous arrangement of the material which forms the crown.

Localities—Middle Fork of Popo Agie River, Bull Lake Creek, Dinwoody Creek, Wind River Mountains; Red Creek, Owl Creek Mountains, Wyoming.

Ancistriodus C. BRANSON, N. GEN.

Tooth high, strongly recurved posteriorly, laterally flattened, serrated on anterior and posterior edges, serrations underlain by dentine tubules.

Ancistriodus multisectus C. BRANSON, N. SP.

This species is represented by nine specimens, which exhibit considerable variability. The holotype is high-crowned, curved backward about 60° above the base, straight near the tip, tapering regularly from the base to the sharp tip, crown laterally compressed above the oval base, anterior and posterior edges of crown serrated. The serrations of the posterior edge are broad, blunt, separated by notches about one-third the width of the serrations, tips of most of the serrations with a slight medial notch, number of serrations on posterior edge 30–40, smaller toward the tip. Anterior edge of crown with about 100 small low serrations, smaller toward tip, few notched.

Base of tooth oval, with large pulp cavity which tapers to a blunt termination near the tip of the crown. The pulp cavity gives rise to small dentine tubules each of which connects with a serration and which comes to a blunt termination beneath the tip of the serration. The tubules are extremely variable within a single tooth. Near the base, the majority are complete and fan out from the central core like the base of a lamp stand. Near the tip, some of the tubules fail to meet the central core, but taper off into wisps, and some are irregular at both ends. At the tip, only a slight lateral fibrous appearance is present. The tubules of the posterior margin are in general much more regular than those of the anterior and may be traced distinctly in the serrations to a point nearer the tip of the crown. These structures are observable in all specimens without sectioning, as the teeth are translucent. The surface of the crown bears fine, indistinct oblique wrinkles from the center of the lateral face outward and upward, which pass into more distinct irregular wrinkles arranged three or four above each serration on the anterior edge.

The holotype is 8.4 mm. high, 2.2 mm. long above the base, and 0.8 mm. wide. The specimen is perfect. Two imperfect specimens are respectively 12.3 mm. and 11.2 mm. high, and a badly broken specimen indicates a somewhat greater size. One specimen is different in that the tip not only is straight but curves slightly anterior above the general posterior curvature of the lower part of the crown.

A. multisectus resembles *Hamatus phosphoriensis* in its claw-like form, and in that *H. phosphoriensis* has a few indistinct serrations near the tip in some specimens and shows a lateral linear arrangement in thin section much like that near the tip of *A. multisectus*. It differs in the possession of branch tubules, distinct serrations throughout, laterally flattened shape, less curvature, and sharp tip. The structure of *Ancistriodus* indicates that each serration developed as an independent unit and that after growth the tubules were cut off from the main core to some extent. There is some suggestion of relationship to members of the conodont group, although of the numerous conodonts in the same bed none are of a similar nature, but are of specialized old-age types.

Localities—Middle Fork of Popo Agie River, Dinwoody Creek, Wind River Mountains, Wyoming.

Cladodus occidentalis LEIDY

This species is represented in the Pustula member by a fragment of the middle cusp of a tooth. There is little doubt of the identity of the tooth, however, as it corresponds in all particulars with typical specimens of the Lower Phosphate member.

Locality—Dinwoody Creek, Wind River Mountains, Wyoming.

Campodus variabilis NEWBERRY AND WORTHEN

A single specimen from which one end is missing represents this species. It agrees with teeth of the lateral series as figured by St. John and Worthen,[4] except that the base extends farther past the crown at the sides of the tooth than in typical specimens. A specimen from the Lower Phosphate referred to *C. variabilis* has regularly spaced crenulations on the crown, differing from typical specimens and from the Pustula member specimen in that regard. The Pustula member specimen is 5.4 mm. high, 3.0 mm. wide, and 15 mm. long in the broken condition.

Locality—Dinwoody Creek, Wind River Mountains, Wyoming.

Helodus subpolitus BRANSON

Typical specimens representing many of the known variations of the species are common in the Pustula member.

Localities—Middle Fork of Popo Agie River, Bull Lake Creek, Dinwoody Creek, Wind River Mountains, Red Creek, Owl Creek Mountains, Wyoming.

Janassa unguicula EASTMAN

A specimen in which the crown is perfect was found in the Pustula member. It is typical in all respects.

Locality—Little Wind River, Wind River Mountains, Wyoming.

Edestus ? SP.

A specimen consisting of the tip of a tooth is triangular, tip rounded, edges sharp with broad-topped, sharp-edged serrations irregular in size, some double, separated by narrow, distinct saddles; surface of crown extensively pitted except on the serrations by the weathering out of fine, evenly distributed punctae. Pulp cavity

[4] O. St. John and A. H. Worthen, "Description of Fossil Fishes," *Geol. Surv. Ill.*, Vol VI, Pl. VIII (1875).

long and narrow, extending nearly to the serrated edges of the crown, punctae communicating with the cavity in regular arrangement. Tooth as preserved 7.5 mm. high, 8.6 mm. long, 3.2 mm. wide. Another specimen is a fragment from near the base of a tooth. Its conformation shows that the first specimen was broken off not far above the base. The crown surface of the second specimen is not eroded, and the punctae do not reach the surface. The surface of the serrations and the adjacent part of the crown is marked by deep, narrow, scratch-like lines which parallel the direction of the serrations.

Locality—Dinwoody Creek, Wind River Mountains, Wyoming.

Ctenacanthus browni BRANSON

A specimen consisting of nearly a foot of the middle section of the spine was found to agree in all particulars with the type. The new specimen was so preserved that the interior cavity is a separate piece. It tapers with the spine in general, but is irregularly constricted and expanded throughout its length, and it is marked by irregularly spaced annular ridges. Many unidentifiable fragments may belong to this species.

Locality—Dinwoody Creek, Wind River Mountains, Wyoming.

Ctenacanthus mutabilis C. BRANSON N. SP.

The species is described from a single specimen, a fragment from near the tip of the spine. Fragment triangular, subcircular in cross-section at tip, long and narrow at base, much flattened laterally in this portion; anterior face rounded, acute, posterior face broad. The tip as preserved is abruptly truncated at the upper anterior corner for the upper one-third of the fragment, but the surface of the truncated area is eroded in such a manner that the truncation may be only apparent. Interior cavity small, flattened elliptical, near posterior margin, terminated below tip of fragment. Posterior face of spine 19 mm. below top of preserved portion with raised shallow trough bounded by distinct shoulders; trough communicating in an abrupt drop with a channel in the posterior face of the lower part of the preserved portion; channel 2 mm. deep at upper end, 6 mm. deep 22 mm. below where fragment ends. Anterior face of spine in middle of fragment with an area of enamel; enamel area broad, rounded,

surface shiny, irregularly folded transversely, distinct median longitudinal groove through lower half of area, area 21 mm. long, 5 mm. wide, may have extended farther toward tip where specimen is eroded, lateral and inferior margins of enamel area abruptly terminated by a tucking in of the enamel surface so that the enamel area stands some 2 mm. above the surface of the spine and is slightly undercut. Surface of spine marked by 15–20 longitudinal ridges to the centimeter, ridges irregular in height and width, with expansions and constrictions, some ridges fade out, others are introduced, and some are oblique to the general parallelism, but do not cross adjacent ridges; ridges near the posterior face of the spine oblique, extending into the trough and channel of the posterior face and forming acute triangles at their inner ends where they approach the normal ridges.

The enamel area seems to have been the only portion of the spine which was not covered by a dermal layer in life. The holotype is 72 mm. high, 49.4 mm. long at the base, 16.8 mm. wide at the base, 21 mm. long at the tip, and 13.5 mm. wide at the tip. The species does not closely resemble any described species of *Ctenacanthus*. It is distinguished by the number and peculiar irregularities of the longitudinal ridges, by the anterior enamel area, and possibly by the truncation of the upper anterior part of the tip.

Locality—Middle Fork of Popo Agie River, Wind River Mountains, Wyoming.

Spine Denticles

Isolated spine denticles are common at all localities. They are cap-like, smooth, and uniform in size, all about 3 mm. in diameter and 2 mm. high. They are like denticles found in the Lower Phosphate and like those also in that none has been found attached to a spine.

Localities—Wind River Mountains and Owl Creek Mountains, Wyoming.

FIG. 1

EXPLANATION OF FIGURE 1

Dolophonodus uncinatus n. sp.

1. Anterior view of holotype. ×$\frac{4}{3}$
2. Posterior view of holotype. ×$\frac{4}{3}$
3. Lateral view of holotype. ×$\frac{4}{3}$
4. Anterior view of paratype. ×$\frac{4}{3}$

Janassa unguicula Eastman

5. Posterior view of crown. ×$\frac{4}{3}$
6. Anterior view of same specimen. ×$\frac{4}{3}$

Campodus variabilis Newberry and Worthen

7. Coronal view. ×$\frac{8}{3}$

Ancistriodus serratus n. sp.

8. Lateral view of holotype. ×6
9. View through anterior edge of paratype. ×70
10. View through posterior edge of paratype. ×40
11. View through crown of paratype. ×10

Hamatus phosphoriensis C. Branson

12. Lateral view of perfect specimen. ×$\frac{8}{3}$

Ctenacanthus mutabilis n. sp.

13. Anterior view of holotype, showing enamel area. ×$\frac{4}{3}$
14. View of right side of holotype. ×1

Edestus ? sp.

15. Lateral view of fragment of crown. ×$\frac{8}{3}$

Cladodus occidentalis Leidy

16. Lateral view of fragment of central cusp. ×$\frac{8}{3}$

STRUCTURAL FEATURES IN THE COLUMBIA RIVER LAVAS OF CENTRAL WASHINGTON[1]

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ABSTRACT

Structures in the Columbia River lavas were studied principally in the Moses Coulee area of central Washington. Here a 1,900-foot section of olivine basalts rests with apparent conformity on Upper Miocene sedimentaries. The lavas exhibit many different types of structure, such as columns of various sizes, cross jointing, irregular fracturing, fanlike forms, rosettes, horseshoe shapes, ellipsoidal and brecciated basalts. The conditions under which the lavas were extruded, the probable methods of formation of the fans, and ellipsoidal and brecciated basalts are discussed.

The Columbia River lavas were studied by the writer at various localities in Washington, Idaho, and Oregon, but most of the specific features referred to in the following article are found in the Moses Coulee area of Washington. Here the basalt section is about 1,900 feet thick and rests with apparent conformity on Upper Miocene[2] sedimentaries. The sedimentary rocks can be readily correlated as equivalent, in part at least, with the Latah and Ellensburg formations, and the overlying lavas with Smith's Wenas basalt.[3]

Throughout the Moses Coulee area many different kinds of basalt structures can be seen. Columns are thick and thin, have parallel walls or wavy outline, and few or numerous horizontal fractures. Many of the flows are not columnar, but massive and broken by widely or closely spaced joints. Some are irregularly fractured throughout and broken into odd-shaped blocks 6 inches to 1 foot in diameter. Several exhibit spheroidal weathering. Still others are largely or entirely ellipsoidal, and a few contain a large percentage of shattered basalt glass.

[1] Published by permission of the supervisor, Division of Geology, Department of Conservation and Development, Washington.

[2] Arnold D. Hoffman, "The Douglas Canyon Flora of East-Central Washington," *Jour. Geol.*, Vol. XL, pp. 735–38.

[3] George Otis Smith, *U.S. Geol. Surv., Ellensburg Fol. 86* (1903), p. 3. Malvin G. Hoffman, "Several Phases of Geology of the Moses Coulee Area," *Jour. Northwest Sci.*, Vol. VI, No. 4 (1932), pp. 120–23.

COLUMNS, FANS, AND ROSETTES

Locally, columnar structure is very well developed, but is not persistent and may within short distances grade into any one of the other types. The diameter of the columns ranges mostly between 1 and 4 feet, and their length from a few feet to 60 or 70 feet. Many of them are approximately hexagonal in outline, but geometric symmetry is not one of their characteristics. Some have parallel sides, but more commonly, especially the larger columns, thicken and thin,

FIG. 1.—Moses Coulee looking northward toward Palisades. The channel here is almost a mile wide and ranges in depth from 600 to 1,100 feet. (M. A. Search.)

the outline being wavy. They are marked by horizontal fractures or fractures normal to the longer axis. These are usually spaced about 2 feet apart, but some are so closely spaced as to give the rock almost a shattered appearance. Where the cross joints are close together they vary mostly between 90° and 60° to the longer axis. Both dense and vesicular columns were seen. In several instances the vesicles are as much as 4 or 5 inches across.

A number of examples of fan arrangement of columns were noted at the base of flows where the lava had ridden over a hill or hump. The columns radiate like the spokes of a wheel, remaining approximately normal to the surface upon which the lava rests, usually

some irregularity on the surface of an underlying flow. The fan structures are not large. In most instances their diameters are only 300 or 400 feet, the columns being 1 or 2 feet across. Another common occurrence of the fans is at the termination of lobes where the surface of the lava flow due to its high viscosity was very abruptly curved. The columns, developing normal to the cooling surfaces, formed fanlike structures 300–500 feet across composed of columns 2–3 feet in diameter.

A few instances of columns arranged in roseate and horseshoe shapes were noted where the lava had filled a neck or channel. Small, poorly developed rosettes were also seen in some of the pillows.

BASALT PATTERNS

In the lower end of Moses Coulee and along the Columbia Canyon for several miles both north and south of the coulee mouth is a flow about 175 feet thick, the base of which is approximately 400 feet above the coulee floor. The top layer is of vertical columns about 4 feet in diameter and of varying lengths from 15 to 40 feet. The bottom layer is approximately the same. Between these two are several lava sheets of smaller columns, 1–2 feet in diameter, which are arranged into many fans and a few horseshoes and rosettes. The flow was not formed as a result of the extrusion of one thick layer, but of many sheets, tongues, and lobes which rose through numerous fissures and came approximately to rest before solidification and jointing. The lava spread out in all directions, the masses approaching, meeting, or overlapping one another and developing a very uneven surface. Though the outpourings occurred in rapid succession, each one to a large extent developed a columnar structure before the next one spread over it. The arrangement of columns in each sheet usually is independent of that in adjacent sheets, and their thinness, 1–2 feet in diameter, suggests relatively rapid cooling. The chaotic pattern resulting from the solidification of the lava on the uneven surfaces is somewhat further enhanced by dike structures. The final extrusion must have been more fluid than the intermediate ones as it smoothed over all of the irregularities, its own surface being quite flat.

The foregoing characteristics were noted in several localities, but

are best seen in the one described. The top layer of vertical columns is not present in all of the flows which possess the erratic fan structures. In such instances a last thick fluid extrusion did not occur.

PROBABLE MANNER OF EXTRUSION

Very few of the flows appear to have resulted from one massive extrusion. Those which from a distance appear to be homogeneous layers 150 or more feet thick are found to consist of a series of sheets

FIG. 2.—Wedging and overlapping of basalt sheets in a single flow. Southeast wall, 1 mile below Three Devils. Height of cliff is about 100 feet.

10–30 feet thick which have merged in some places, but have retained their identity in others. The merging, pinching out, and replacement of the sheets occur within such short distances it becomes apparent that the lava did not travel far from its source. Lava in the very thin layers may have moved but a few hundred yards from the point or line of eruption while in the thicker layers in most instances it probably did not flow more than a few miles. There is nothing to indicate that the same passages were used by successive eruptions. Where dikes were seen they could not be traced to more than one flow. Although some of the old fractures may have been reopened, many new fissures were undoubtedly made immediately preceding each extrusion.

The basic magma was probably intruded close to the surface and initiated a condition in which the overlying crust floated on the liquid substratum. Tension fractures were produced by the uneven sinking of this crustal layer. The liquid was forced to the surface largely by the pressure of the overlying rock, and to some degree by the force back of the intrusion. Each outpouring occurred simultaneously with the sinking of the land surface. The latter change occurred periodically on account of the tendency of the solid not to break until after its elastic limit has been reached. In this manner further settling and fracturing took place after the extrusion and solidification of each flow. The process continued until all, or nearly all, of the lava had been forced to the surface, the final surface level probably being not much changed from what it was at the start.

ELLIPSOIDAL BASALTS AND BASALT BRECCIAS

The lower half of the basalt section in Moses Coulee is largely ellipsoidal, brecciated, and scoriaceous. The thickness of each type varies considerably from place to place. The scoriaceous portion is usually at the base or top of a flow and commonly 5–20 feet thick, the greater thickness on top. The brecciated flows are ordinarily 50–75 feet thick, and the pillow basalts about the same. In a few places three or more ellipsoidal layers rest upon one another, showing a thickness locally of at least 250 feet.

All structural types grade into one another, some gradually and some abruptly. Here and there a lens of columnar lava is included within ellipsoidal or scoriaceous basalt. In a few places ellipsoidal lava grades upward into columnar basalt.

Four structural types were found which are gradational from ellipsoidal basalts to basalt breccias. The first is composed of pillows with open spaces between them. Most of the pillows are from 2 to 6 feet long and from 1 to 3 feet thick. Occasionally are seen large pillowy masses 15–20 feet across. They have a 1- or 2-inch dense black, glassy coat within which is either dense or vesicular rock. Internally some of them possess a roseate structure. The spaces between the bulbous masses are 1 or 2 feet across and 2 to 4 feet long, the sidewalls of which are curved inward.

The second type is like the first except that the spaces are filled

with a brecciated basaltic glass. The breccia is yellowish in appearance, owing to the yellow coating of the weathered glass.[4] In a number of instances the lower part of a flow is of the first type, which grades upward into the second.

In the third type the pillows are separate and completely surrounded by yellow-coated brecciated glass usually 1 or 2 feet thick.

FIG. 3.—Ellipsoidal lava grading upward into a phase in which the spaces between the pillows are filled with brecciated basalt and overlain by columnar basalt. Near the base of the flow some large pillows are noticeable. Height of exposed cliff about 110 feet.

These pillows do not have the characteristic droplike shape. Their cross sections are ellipses, some of them very nearly circles.

The fourth type is largely breccia with stringers and small ellipsoidal masses of basalt distributed through it. At a number of places the stringers and rows of pillows are arranged more or less parallel to each other and give the flow a rude, cross-bedded appearance. The stringers are 3–5 feet thick while the interbedded layers of breccia are commonly 5–15 feet thick. Maximum dips of this apparent cross bedding are as high as 30°. A few showed dips as low as 9° or 10°. An example of the latter is located at the mouth of Douglas

[4] Martin A. Peacock and Richard E. Fuller, "Chlorophaeite, Sideromelane and Palagonite from Columbia River Plateau, *Amer. Mineralogist*, Vol. XIII (1928), p. 375.

Canyon. The dip now measures 15°, but tilting subsequent to extrusion accounts for 6° or 7°.

Lewis[5] has stated that ellipsoidal structure may be developed under either subaerial or subaqueous conditions, which seems to be borne out by field evidence. He further stated that the pillows undergo no movement after formation and that the breccia filling is

FIG. 4.—Ellipsoid showing a crude roseate structure surrounded by basalt breccia. Longest dimension of pillow is 8 feet. (M. A. Search.)

due to spalling from their surfaces. The latter statement is not entirely in accord with some of the writer's observations. In the third type described above the pillows are separate and completely surrounded by breccia often a foot or more in thickness. Some rolling motion of the pillows is required to develop this kind of structure.

The flows which are largely breccia and which are marked by apparent cross bedding, according to Fuller,[6] have been formed as follows:

> A fluid lava on encountering a body of water would tend to granulate like molten slag. . . . . The fine breccia would settle until its surface attained an

[5] J. Volney Lewis, "Origin of Pillow Lavas," *Geol. Soc. Amer.*, Vol. XXV (1914), p. 652.

[6] R. E. Fuller, "Aqueous Chilling of Basaltic Lava on the Columbia River Plateau," *Amer. Jour. Sci.*, Vol. XXI (1931), p. 284.

angle of repose which, owing to the roughness of the fragments would be relatively steep. . . . . The accumulation of granulated glass would gradually advance like the foreset bedding of a delta. The inclined bedding would be preserved by the thin sheets and the ropy or ellipsoidal masses, which failed to granulate.

If the fine breccia settled in water, it would show some sort of stratification. The breccia layers in the "foreset bedded" flows are not in the least stratified. The apparent bedding is due only to the alinement of the sheets, stringers, and ellipsoidal masses.

Fuller[7] further assumes that all the ellipsoidal lavas in the region were formed under water, upon which he bases his statement that lakes existed at the time of their extrusion. This appears to be founded on the idea that pillow structure can only be developed when the lava is quickly quenched as a result of being poured into water. Neither the studies by Lewis[8] nor the field evidence substantiate this assumption. The Douglas Canyon sediments are distinctly subaqueous and are directly overlain by basalt which is not ellipsoidal. Fuller remarks the absence of pillow structure at this locality even though he attempts qualification by stating that a suggestion of ellipsoidal structure exists. This is the only locality in the Moses Coulee area where there is some basis for the implication that a body of water of some sort probably existed at the time of the extrusion of the lava.

A few observed instances of the accumulation of pillows in water have perhaps been the reason why so much weight is usually placed on this mode of formation. Anderson[9] witnessed their formation in a thin sheet of basalt flowing into the ocean on the coast of Savaii. Green[10] described a similar occurrence on the northeastern coast of Hawaii:

The red-hot molten lava was quietly tumbling into the sea over a ledge perhaps 6 to 8 feet high and 500 to 600 feet long. The lava did not seem to be quite so liquid or of such bright color as it did when it ran out of the openings in the

[7] *Ibid.*

[8] *Op. cit.*

[9] Tempest Anderson, "Volcano of Matavanu in Savaii," *Quarterly Jour. Geol. Soc.*, Vol. LXVI (1910), pp. 631–40.

[10] W. L. Green, "Vestiges of the Molten Globe, Part II: "The Earth's Surface Features and Volcanic Phenomena" (Honolulu, 1887), pp. 172–73.

side wall of the aa stream up in the mountains some months before. It ran more like porridge in great flattened spheroids, which were sometimes partially united together and sometimes almost separate.

Green here indicates that the spheroidal masses were formed before they entered the water.

Lava that is very viscous and near the solidification temperature at the time of extrusion would pile up into a large heap near the

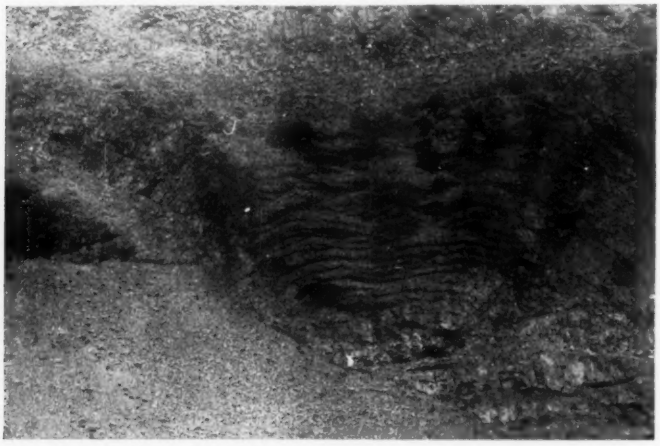

FIG. 5.—Wrinkles in lava on steep slope at the end of a lobe. The wrinkles are about 4–6 feet wide. East wall Columbia gorge, ¾ mile north of Moses Coulee.

vent, the size depending on the amount of material and the pressure back of it. If the same flow were very hot when it reached the surface, it would be quite fluid and spread out into a rather even layer. Columnar structure would result if the lava came to rest before solidifying and was subjected to little or no disturbance during cooling. The formation of the ellipsoidal and brecciated basalts of central Washington requires the supposition of further conditions, and, according to the writer's interpretation, originated in the following manner. The lava was extruded in a series of more or less rapidly flowing thin sheets over a rough and uneven surface. The rate necessary was such that at the time of the formation of the pillows most of the mass was in motion. The temperature of the eruption may,

or may not have been much above the solidification point. If the liquid were quite hot the ellipsoids would have been formed farther away from the source than if the temperature were close to the freezing point. As the lava cooled and its viscosity increased any rounded shapes which it assumed would be retained. The upper layers of the lava moved faster than the lower layers. As the liquid rock passed over the humps those small areas of the surface immediately beyond moved down slopes which dipped at angles greater than that of the general surface inclination of the flow. This increased slope gave these small areas a velocity higher than that of the general surface rate of motion. If the velocity difference was great enough and the lava sufficiently liquid, the small masses would overrun not only the lower layers of the flow, but also the downgrade adjacent portions of the surface. The resulting shapes would be ellipsoidal forms of various sorts, most of them resembling miniature recumbent folds. If a crust had already formed it might be fractured when folded over and through these fractures would be forced small tongues of lava, the larger rounded ends pointing in the direction of the flow. Fractures without folds might have been produced in the crust by the stretching of the surface during its passage over the humps. This would also allow for the development of ellipsoids by the extrusion, through these openings, of lava in small bulbous masses. If immediately after the formation of each ellipsoid it came to rest, relative to the others with which it was in contact, type No. 1 would be developed with cavities between the pillows. In this manner pillow structure could be formed simultaneously over a large area. A considerable thickness of ellipsoidal basalts would be produced by a series of such extrusions, the rounded, hummocky surface of each aiding in initiating the budding of each succeeding layer.

This mode of formation is suggested by the presence of corrugations and humps found occasionally on basalts, especially so at the ends of lobes where the surface is rather steep. The width of the irregularities varies in size from a quarter of an inch to several feet.

If the surficial crust had cooled until it was quite brittle, a great deal of spalling would occur during the folding over of the lava into ellipsoidal masses, the glassy breccia filling the spaces between the pillows, forming type No. 2.

Those pillows which were formed before a solid crust made its appearance would tend to stick together while those formed after the surface had solidified could be easily parted. An increased velocity under the latter condition would completely separate the pillows, each one moving forward in a sort of rolling motion. There would be a considerable amount of spalling, the projecting necks being broken off first. The farther the elliptical masses rolled, the more rounded they would become and the greater the thickness of basalt breccia between them. This would develop the third type described above. The distance the individual pillows may have rolled could very probably be measured in scores of feet, as is suggested by the method of development of the fourth type.

To explain the apparent cross bedding in the brecciated flows (fourth type) requires a further expansion of the idea described above. The thin sheets of lava must have been extruded at such a velocity that the momentum of the mass was sufficient to overcome the resistance to shearing within the layers as the lava cooled to a solid, shattering most or all of the body. Some pillows would be formed, and rolling would reduce their size. If several were formed at one time they might come to rest before being completely reduced to breccia and thus would mark the top of one of the layers. Some of the pillows may have been quite large at the start. If one of these broke in transit the contained liquid would pour out forming a stringer of lava. In this way layers of unsorted breccia would be marked by layers of small pillows and lava stringers, giving the whole an appearance of bedding. Thus the low-angle "stratification" would be formed. To develop the high-angle "cross bedding" it is necessary to carry each successive sheet of lava forward until, nearly or entirely solidified, it was pushed over the end of the flow. The shattered rock would roll down the slope at the end of the lava flow, coming to rest at an angle of repose of approximately 30°. The entire extrusion with a thickness of 50 or 75 feet would be advanced by the addition of a series of steeply dipping layers. In this manner it could be carried forward for a quarter to half-mile, as is seen at several places in the Coulee region.

PETROGRAPHY

The lavas in the Moses Coulee area are all basalts which vary in color from gray and dark reddish-brown to black. In the whole 1,900-foot section the principal distinguishing characteristic is texture. There are all gradations from glass to nearly holocrystalline rock. Even those specimens which are medium grained contain some dense grayish-brown, and/or reddish-brown, and/or bottle-green glass and cryptocrystalline material in the intercrystal spaces and occasionally as cavity fillings. Many of the gas holes in the vesicular rocks are partially or completely filled with opal, calcite, and zeolites. In much of the breccia the glass totaled 85 per cent, while, in all the other samples it was considerably less, the minimum being about 1 per cent. The texture of the medium-grained rocks is ophitic, and that of the fine-grained to glassy specimens is intergranular, felty, or porphyritic.

An average mode is as follows: labradorite 52 per cent, augite 39 per cent, olivine 4 per cent, magnetite 3 per cent, cryptocrystalline material and glass 2 per cent. All the minerals are quite fresh and unaltered. The labradorite occurs in needles and laths. Many of the augite crystals have a small $2V$-angle, approximately 30°, and may probably be pigeonite. Most of the olivine is present in small irregular grains. The outlines of the medium-sized olivine crystals show distinct resorption effects. The magnetite is present in small crystals, grains and dendritic forms in the intercrystal spaces, and appears to have been the last mineral to crystallize. No accessory minerals were noted.

STUDIES FOR STUDENTS

ORIGIN OF THE IGNEOUS ROCKS OF MINNESOTA

FRANK F. GROUT

ABSTRACT

The final reports of the Minnesota Geological and Natural History Survey thirty years ago contained several suggestions that igneous magmas in Minnesota were formed almost in place by hydrothermal fusion of earlier rocks. The suggestions did not have the positiveness of conclusions because there were not sufficient chemical analyses to check the compositions. Such analyses are now available and are here given. They do not to any extent support the early suggestion.

Suggestions made elsewhere of such fusion of rocks in the earth's crust are briefly noted and an attempt is made to show about how far the idea of fusion in the crust can now be safely applied.

INTRODUCTION

About forty years ago the geologists of the Minnesota Geological Survey and some members of the United States Geological Survey became convinced that certain intrusive masses in Minnesota originated from earlier sediments and volcanics by a process of fusion, more or less hydrothermal in character, probably not far from the surroundings in which they are now found. Of the three most important petrological conclusions from years of study, H. N. Winchell reported that one was the origin of the great masses of granite and a second was the origin of the Duluth gabbro, by fusion, essentially in the places they now occupy (see Fig. 1).[1] The idea of such a fusion is clearly not a popular one in this day, but these early suggestions of fusion have had little critical discussion. As the rocks are about the only ones in the United States for which such an origin has been strongly maintained and not much criticized, it is clearly in order to re-examine briefly the theory in the light of the great mass of data recently accumulated. The discussion may appropriately be conducted by geologists of the same survey that Professor

[1] N. H. Winchell, *Minn. Geol. and Nat. Hist. Surv. 17th Ann. Rept.* (1899), pp. 24, 28, 29, 30, and 36; *Final Rept.*, Vol. V (1900), p. 972. See also *Jour. Geol.*, Vol. XI (1903), p. 131.

Winchell started. The details will be followed by a short general discussion of the depth of rock fusion.

The new data are chiefly the chemical analyses of the several formations. These have been made partly for the survey, partly in

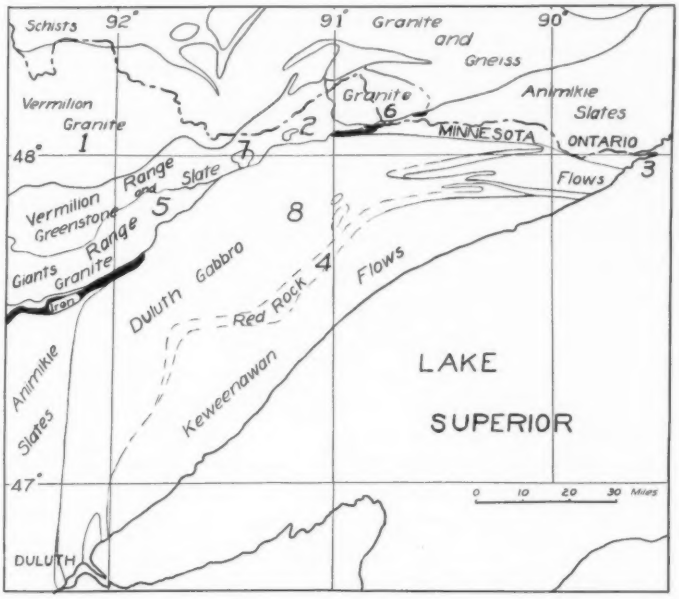

FIG. 1.—Key map showing the geologic setting of the igneous rocks in northeastern Minnesota. Several have been attributed to fusion in place.

1. Vermilion granite
2. Kekequabic granite
3. Pigeon Point
4. Red rock of the Duluth gabbro
5. Giants Range granite
6. Saganaga granite
7. Snowbank syenite
8. Duluth gabbro

personal researches, and partly by student assistants working under a grant from the research funds of the graduate school of the University of Minnesota. All this aid in the study is gratefully acknowledged. It may be added as a tribute to the insight of the older geologists that they realized that the final test of their theory was to be made by chemical work that they were not yet able to supply.

Winchell keenly referred[2] to "the excess of silica and potash that characterize the alkaline magma," and says: "The question is one that is chemical . . . . and leads to an examination of . . . . potassium and . . . . its probable activities under Archean conditions. . . . ." Van Hise[3] notes that a fused shale should possess the essential composition of a sedimentary rock even though fusion made the textures igneous. He notes specifically the need for analyses to settle the origin of two Minnesota masses.

The use of analyses in testing the hypotheses of fusion *in situ* can hardly be questioned. Nearly everyone who has discussed the problem has realized that if igneous rocks are fused in place they should have a composition that is closely determined by the composition of the formations that were there first. Only minor changes due to reaction or addition can be admitted.

VERMILION GRANITE

The Vermilion granite occupies most of the northern portion of St. Louis County, Minnesota. The northern contact phases were described by Lawson[4] and the southern by Winchell with similar conclusions.

Winchell published in 1897 his conclusions on the transitions from mica schist to granite on Outlet Bay of Lake Vermilion (Sec. 14, T. 63 N., R. 17 W.) after having examined it in 1887, and again in 1896 with a party of other geologists. He says[5] the field evidences prove "that the Laurentian igneous rocks are due to fusion of the crystalline schists, and further that the same process which produced the schist from sedimentary strata, produced also when carried to the end result, the igneous rock." He listed under fourteen heads the features not compatible with more remote igneous origin; covering chiefly the *lit-par-lit* injection of granite, pegmatite, and quartz. He referred in one place to a "silicification" as accompanying the

[2] *Amer. Geologist*, Vol. XXII (1898), pp. 309–10.

[3] C. R. van Hise, "Treatise on Metamorphism," *U.S. Geol. Surv. Mono. 47*, pp. 730–36.

[4] A. C. Lawson, "Geology of the Rainy Lake Region," *Ann. Rept. of the Geol. and Nat. Surv. Can.* (1888), p. 131F.

[5] "New Features in the Geology of Minnesota," *Amer. Geologist*, Vol. XX (1897), p. 41.

process, but evidently found no reason to assume great changes in any constituent.

The phenomena of Outlet Bay would now be explained by most geologists as due to granite intrusion at a depth where *lit-par-lit*

TABLE I

ANALYSES OF VERMILION GRANITE AND ADJACENT MICA SCHIST

	Vermilion Batholith				Mica Schist		
	1	2	3	4	5	6	7
SiO_2	58.55	71.73	69.01	72.06	63.04	64.48	63.01
Al_2O_3	12.56	14.76	16.92	16.00	16.45	14.83	16.40
Fe_2O_3	1.72	.58	.55	.46	1.32	1.49	1.85
FeO	5.05	1.35	1.43	.72	4.09	2.40	4.86
MgO	8.63	.62	1.28	.97	5.04	2.32	3.39
CaO	7.34	1.18	2.67	.86	3.17	8.42	3.61
Na_2O	3.09	3.58	4.59	4.56	2.62	2.79	2.44
K_2O	2.01	4.63	2.55	3.54	2.14	1.43	2.40
H_2O+	.50	.64	.41	.39	.66	1.00	1.15
H_2O-	.12	.20	Trace	.05	.03	None	.15
CO_2	None	N.d.	N.d.	.10	N.d.	.70	.27
TiO_2	.51	.53	.23	.12	.54	.35	.56
ZrO_2	.03	.06	.04	.03	.26	.....	.....
P_2O_5	.14	.14	.11	.09	Trace	.....	.....
FeS_2	.04	.06	(S) .04	(FeS_2) .09	.....	.....	.....
Cr_2O_3	.09	.02	.01	.02	.....	.....	.....
MnO	.17	.03	.03	.06	.22	.....	.....
SrO	.02	N.d.	N.d.	None	.....	.....	.....
BaO	.08	.14	.03	.12	.....	.....	.....
	100.65	100.25	99.90	100.24	100.38	100.21	100.09
Specific gravity	2.820	2.615	2.670	2.637	2.735	.....	.....

1. Border phase of Vermilion granite, near Vermilion Lake, Sec. 30, T. 63 N., R. 16 W. F. F. GROUT, analyst.
2. Vermilion granite, Sand Point Lake. F. F. GROUT, analyst.
3. Dark phase of Vermilion granite, that probably assimilated some biotite schist. F. F. GROUT, analyst.
4. Vermilion granite, near Pelican Lake. F. F. GROUT, analyst.
5. Biotite schist, near the dam at outlet of Vermilion Lake. S. DARLING and F. GROUT, analysts.
6. Biotite schist, south of Tower. W. YARWOOD, analyst.
7. Biotite schist, Rainy Lake. Average of three analyses on three beds in a hand specimen; about one-inch beds, ranging from 56 to 76 per cent SiO_2. GROUT and YARWOOD, analysts.

injection and thrusting of walls was about to give way to stoping, both processes being clearly active. It is noteworthy, however, that the analyses show clearly the impossibility of formation of the granite by simple fusion of the schist or any similar sediment. Table I

includes analyses of several phases of the granite, the most abundant, 4, placed beside that of the most abundant type of schist, 5. The contrast is great, especially in soda and iron. The schist seems to be derived from Knife Lake sediments resembling those analyzed for Table II, Analyses 1–4.

It is interesting to note that the granite has not only intruded the schist, but may have assimilated a little. The main granite has little mica, but some exposures near the schist seem to be more micaceous than average. Analysis 3 is of such a phase, and a calculation suggests that possibly 25 per cent of schist has been assimilated. This, however, is not in any sense suggestive of the origin of the granite; it cannot be believed that a batholith 30 by 80 miles across can be so enriched in silica and alkalies as to convert the composition from Analysis 5 into Analysis 4.

On the contrary, the theory of differentiation now favored indicates that a primary magma of even basaltic composition may leave a granitic upper phase in large volumes. Such a history for this granite is indicated by the basic border phase occurring at several points (Analysis 1).

KEKEQUABIC GRANITE

A few months after issuing his argument on the Vermilion granite, Winchell presented a similar discussion of the Kekequabic granite, with the "conclusion that the same rock is both clastic and intrusive and therefore as an intrusive it is derived from the clastic beds *in situ*, and had no deep seated source." Both N. H. Winchell and A. Winchell concluded that there was "partial or complete aqueo-igneous fusion of the sedimentaries of the region," enough to allow it to be intruded into the surrounding rocks, but this intrusion was very local.[6] The evidence is largely in the occurrence of twinned and zoned feldspar, and of alkali pyroxene, in both clastic rocks and granite. Structural features were also cited. "(1) Presence of ancient but partially obliterated lines of sedimentation in the granite. (2) Presence of pebble forms in the granite. (3) Transition from the granite to the clastic rocks. (4) Presence in the immediate vicinity of altered clastic rocks resembling the granite." The supposed sedi-

[6] *Ibid.*, Vol. XXII (1898), pp. 299–310.

ments, altered to resemble a porphyry, were named "porphyrel" by Alexander Winchell.[7]

Dr. U. S. Grant dissented from the conclusions[8] on the basis of later detailed field and microscopic study. The first three field struc-

TABLE II

ANALYSES OF KNIFE LAKE SLATE AND KEKEQUABIC GRANITE

	Slate and Graywacke				Granite	
	1	2	3	4	5	6
SiO_2	54.71	61.39	64.05	64.37	66.84	67.42
Al_2O_3	20.52	16.97	15.14	13.09	18.22	15.88
Fe_2O_3	1.72	.39	1.07	.36	2.27	1.37
FeO	6.40	5.32	4.00	4.32	.20	1.14
MgO	4.76	3.84	2.10	2.98	.81	1.43
CaO	1.93	3.21	3.65	3.49	3.31	3.49
Na_2O	2.83	2.78	3.34	3.26	5.14	6.42
K_2O	2.68	1.25	3.47	4.33	2.80	2.65
H_2O+	3.25	2.44	1.60	1.16	.46	.05
H_2O-	.15	.06	None	.08	.....	.....
CO_2	.....	.88	.....	.....	.....	.....
TiO_2	.89	.62	.46	.34	.....	.....
ZrO_2	.....	.07	.....	.....	.....	.....
P_2O_5	.....	.19	.....	.....	Trace	.07
S	.....	.15	.11	.....	.....	.....
Cr_2O_3	.....	.01	.....	.....	.....	.....
MnO	.....	.12	.....	.....	.....	Trace
Others	.....	.06	1.70	1.92	.....	.....
	99.84	99.75	100.69	99.70	100.05	99.92
Specific gravity	2.81	.....	.....	.....	.....	.....

1. Knife Lake slate, from type locality near the west end of Knife Lake, Minn. GROUT and HARTWELL, analysts.
2. Graywacke slate, a coarse phase of Knife Lake slate, from the type locality. F. F. GROUT, analyst.
3. Knife Lake slate near Jordan Lake, Minn. W. S. YARWOOD, analyst.
4. Graywacke phase of Knife Lake slate, near Jordan Lake, Minn. W. S. YARWOOD, analyst.
5. Augite soda granite typical of the stock at Kekequabic Lake, Minn. U. S. GRANT, analyst.
6. Granite porphyry, Kekequabic Lake. U. S. GRANT, analyst.

tures mentioned could not be found, and the fourth is not at all conclusive evidence of origin, since the altered clastics referred to are not in position intermediate between the granite and other clastics.

[7] *Minn. Geol. and Nat. Hist. Surv. 15th Ann. Rept.* (1887), pp. 158–59.

[8] "The Geology of Kekequabic Lake," *21st Ann. Rept. Geol. and Nat. Hist. Surv.* (1893), pp. 5–58, and *24th Ann. Rept.* (1899), p. 104.

The contact rocks in this district are more varied than in the Vermilion granite, partly Knife Lake slates, partly Ogishke conglomerate, and partly a tuffaceous rock that is in close relation to these formations.[9] The composition of the Ogishke conglomerate is too variable to be shown in analyses; the tuff has not been analyzed, but consists so largely of amphibole that it can hardly be a source of granite; the analyses of Knife Lake slate are here given for comparison with that of granite long ago carefully studied by Dr. Grant. The granite has much more soda and much less iron oxide and magnesia than the slates.

RED ROCK OF PIGEON POINT

The eruptive rocks of Pigeon Point at the northeast extremity of Minnesota were selected by the United States Geological Survey geologists as types of eruptives of the Lake Superior Huronian and Keweenawan areas, and W. S. Bayley in 1893 reported[10] his conclusion that the red rock, keratophyr or granite, was in all probability "produced by the fusion of the slates and quartzites of the Animikie through the action upon them of the gabbro. The magma thus formed acted in all respects like any intrusive magma."

In making his detailed study of Pigeon Point, Bayley did not neglect to check his conclusions by chemical analyses, several of which are given below. He called attention to the fact[11] that the red rock and average sediment differ considerably in iron and alkali content, but he assumed that the "true average" of sediments contained more alkali than his average analysis. Bayley himself in another place called attention to an exposure[12] where the uncertainty of the composition of the sediment can be eliminated, viz., a red-rock rim around a quartzite inclusion in gabbro. If this red-rock rim was formed by fusion of quartzite, it was formed by fusing this particular quartzite. Bayley says of the red-rock rim that the circumstances "seem to prove its origin to have been due to the fusion

[9] Mapped in recent years by Dr. U. S. Grant, of Northwestern University, and as yet unpublished.

[10] "The Eruptive and Sedimentary Rocks of Pigeon Point, Minnesota," *U.S. Geol. Surv. Bull. 109* (1893).

[11] *Ibid.*, pp. 112–13.

[12] *Ibid.*, pp. 102, 110, 116.

of the experior portion of the inclusion." He did not, however, have this particular quartzite and red rock analyzed. The exposure was so accurately described that it was possible to go directly to the place without waste of time and samples have been analyzed by Dr. T. M. Broderick. The differences detected in Bayley's analyses of the whole formations are of the same order as the differences between the quartzite inclusion and its red rim. The red-rock rim has 8.33

TABLE III*

PIGEON POINT ROCKS

	GABBRO	INTERMEDIATE ROCK	RED ROCK	FRESH QUARTZITE	FRESH SLATE	ESTIMATED AVERAGE SEDIMENT
	1	2	3	4	5	6
SiO_2	49.88	57.98	72.42	73.85	59.71	70.31
Al_2O_3	18.55	13.58	13.04	10 91	18.32	12.81
Fe_2O_3	2.06	3.11	.68	6.98	8.11	7.26
FeO	8.37	8.68	2.49	.89	.85	.88
MgO	5.77	2.87	.58	1.52	3.54	2.03
CaO	9.70	2.01	.66	.44	1.05	.60
Na_2O	2.59	3.56	3.44	2.28	1.93	2.19
K_2O	.68	3.44	4.97	1.39	3.43	1.90
H_2O	1.04	2.47	1.21	1.88	3.24	2.22
Etc	1.46	2.21	.84	.05	Trace	Trace
	100.21	99.91	100.37	100.19	100.18	100.20

* All analyses quoted from W. S. Bayley, *U.S. Geol. Surv. Bull. 109.*

1. Composite sample of gabbro, Pigeon Point (p. 37).
2. Intermediate rock, Pigeon Point (p. 63).
3. Composite of granular rock (p. 56).
4. Average of three analyses unaltered quartzite, Pigeon Point (p. 84).
5. Very slightly altered slate, Pigeon Point (p. 84).
6. Average of the three quartzites and one slate, Pigeon Point (p. 113).

per cent alkali, while the gabbro has only 3.49 per cent and the quartzite 3.60 per cent alkali. No fusion of a quartzite with 4 per cent alkali could raise the alkali content to 8 per cent, and no assimilation of such quartzite in gabbro with 4 per cent alkali could give a hybrid with 8 per cent alkali. These processes, as far as they may have operated, were overshadowed by a differentiation of the magma. Compare the analyses of Table III, especially column 3 with column 6.

Professor R. A. Daly has twice written regarding the origin of the

red granite of Pigeon Point.[13] He was at first inclined to accept Bayley's conclusion, but in view of the chemical evidence, he now believes with the geologists of this survey that "as the proximate mode of origin for the red rock, differentiation must be assumed." Back of that process Daly calls attention to the many possible and probable loci of assimilation. The evidences of some slight marginal assimilation seem to be accepted by everyone who has seen the field or microscopic sections, but the importance and extent of more remote action may be questioned. The proportion of red rock in the sill at Pigeon Point is not much greater than could be derived by differentiation from average gabbro, viz., about 1 to 7;[14] and, on the other hand, the addition of average Animikie sediment to any rock exposed on Pigeon Point to make as much as 30 per cent of the mixture would result in a rock with conspicuously hybrid-looking analysis.[15] Reversing the calculation and subtracting sediment from the red rock also result in an anomalous residue, if the amounts of assimilated sediment are assumed (for the test) to be above 30 per cent of the red rock. Since no evidence appears to emphasize the importance of assimilation, it is probably safer to estimate somewhere near halfway between the insignificant amount visible and the 30 per cent which seems to be the upper limit based on calculations.

MINOR EXAMPLES OF GRANITE ATTRIBUTED TO THE FUSION OF OLDER ROCKS

Winchell described the "red rock" associated with the Duluth gabbro as fused sediment,[16] even referring to the pink Giants Range granite as the extreme exhibition of the outburst of red rock.[17] As to the Giants Range rock, he suggested[18] that the greenstone was

[13] "The Secondary Origin of Certain Granites," *Amer. Jour. Sci.*, Vol. XX (1905), p. 185; "The Geology of Pigeon Point, Minnesota," *ibid.*, XLIII (1917), p. 423.

[14] Frank F. Grout, "The Use of Calculations in Petrology," *Jour. Geol.*, Vol. XXXIV, (1926), p. 551.

[15] Gabbros with over 10 per cent quartz are known, but make up only about 1 in 500 (*U.S. Geol. Surv. Prof. Paper 99*).

[16] N. H. Winchell, *Minn. Geol. and Nat. Hist. Surv. 10th Ann. Rept.* (1882), pp. 112–13, and *15th Ann. Rept.* (1887), p. 352.

[17] *Minn. Geol. and Nat. Hist. Surv. 15th Ann. Rept.* (1887), pp. 348–55, and *16th Ann. Rept.* (1888), p. 104.

[18] *Minn. Geol. and Nat. Hist. Surv. Final Rept.*, Vol. IV, p. 276.

gradually converted into red granite. The analyses of Table IV show that this is wholly improbable. Furthermore, since the Pigeon Point

TABLE IV

ANALYSES OF ELY GREENSTONE AND NEARBY GRANITES

	Greenstone		Giants Range Granite		Saganaga Granite	Snowbank Granite
	1	2	3	4	5	6
SiO_2	52.46	51.95	66.31	71.45	68.81	69.50
Al_2O_3	16.77	12.58	12.70	12.99	16.30	16.86
Fe_2O_3	3.92	.90	1.98	2.44	1.95	.50
FeO	5.44	8.77	2.66	1.78	.75	.64
MgO	5.20	8.90	3.01	.84	.91	1.45
CaO	6.23	7.00	4.77	.65	2.19	1.70
Na_2O	2.58	2.79	5.03	3.03	4.97	4.58
K_2O	1.39	1.38	2.33	4.79	2.29	3.94
H_2O+	3.28	2.67	.75	1.24	1.38	.42
H_2O-		.14	.05	.10	.12	.20
CO_2	2.14	1.02	N.d.	N.d.	.30	.....
TiO_2	.45	1.03	.49	.13	.22	.53
ZrO_2	.....	.....	.06	.10	.02	.....
P_2O_5	.16	.....	.13	.04	.07	.....
S	.....	.11	.....	.....	.03	.....
Cr_2O_3	.....	.17	.01	.02	.02	.....
BaO	.....	.....	.....	.....	.10	.....
MnO	.....	.15	.04	.04	.02	.....
	100.02	99.56	100.32	99.64	100.45	100.32
Specific gravity	.....	.....	.....	.....	2.657	.....

1. Average greenstone, estimated from many analyses by CARL ZAPFFE. See *Econ. Geol.*, Vol. VII (1912), p. 149.
2. Composite sample of greenstone, Vermilion Lake, Minn. S. DARLING, analyst. See also several similar analyses reported by G. M. Schwartz, *Jour. Geol.*, Vol. XXXII (1924), p. 100.
3. Hornblende-biotite-granite, Giants Range, Minn. I. S. ALLISON, analyst. See *ibid.*, Vol. XXXIII (1925), p. 505.
4. Biotite-granite, Giants Range, Minn. R. J. LEONARD, analyst. Same reference, and see also *Geol. Soc. Amer. Bull.*, Vol. XL (1929), p. 805 n.
5. Saganaga granite, Saganaga Lake, Minn. F. F. GROUT and A. J. BAUERNSCHMIDT, analysts.
6. Snowbank granite. C. W. SANDERS, JR., analyst. See *Jour. Geol.*, Vol. XXXVII (1929), p. 143.

red rock is a differentiate of gabbro with relatively little effect by the assimilation of sediments, the red rock of the gabbro area is also probably a differentiate.[19] This is the more conclusive since the roof

[19] Grout, "A Type of Igneous Differentiation," *op. cit.*, Vol. XXVI (1918), p. 636.

of the gabbro is in most places part of the series of Keweenawan lava flows, not a sediment; the floor also is not so silicic as the rocks of Pigeon Point (see the analyses contrasted with gabbro in Table V).

The relations of the Saganaga granite were fairly well settled in the later reports of the survey; the basal conglomerate of the overlying Ogishke formation is a "recomposed granite" so much like the older granite that in early work it was mistaken for it in the field.[20] Alexander Winchell first described the "syenite" as both massive and forming the matrix of a conglomerate, the matrix having been rendered plastic by "igneo-aqueous" action.[21] H. V. Winchell also maintained the development of the "syenite" by metamorphism of the Keewatin greenstone and associated rocks.[22] Grant, however, sampled the granite and transitions in 1893, and his study of the thin sections left no doubt as to the distinctness of the igneous and sedimentary rocks. His work was reviewed by a party later and the deep-seated origin of the granite settled without need of analytical work,[23] though analyses are given in the final reports showing ordinary granite very different from the greenstone and other rocks adjoining. New analyses show more alkali than the old analyses, but the rock is still to be classed as "sodic granite."

Some quartz porphyries west of Snowbank Lake were described by Winchell with the suggestion that they were formed by alteration of an oceanic precipitate.[24] These were mapped by Clements three years later as normal formations of the Vermilion Range.[25] Winchell's conclusions are said to have been influenced by the similarity of granite at this lake to that at Kekequabic. He includes the granite (syenite) as a product of fusion of fragmental rock. Table IV

[20] N. H. Winchell, *Minn. Geol. and Nat. Hist. Surv. Final Rept.*, Vol. V, pp. 16–17.

[21] *Minn. Geol. and Nat. Hist. Surv. 16th Ann. Rept.* (1889), p. 219, and "Conglomerates in Gneissic Terranes," *Amer. Geologist*, Vol. III (1889), p. 158.

[22] H. V. Winchell, "Geologic Age of the Saganaga Granite," *Amer. Jour. Sci.*, Vol. XLI (1891), p. 389.

[23] U. S. Grant, *Minn. Geol. and Nat. Hist. Surv. 20th Ann. Rept.* (1893), p. 92; see also *24th Ann. Rept.* (1899), pp. 20–23.

[24] N. H. Winchell, *Minn. Geol. and Nat. Hist. Surv. Final Rept.*, Vol. V, pp. 41, 976–78; Vol. IV, p. 288.

[25] J. M. Clements, "The Vermilion Iron Bearing District," *U.S. Geol. Surv. Mono. 45* (1903), Pl. II.

includes a careful analysis of the syenite for comparison with those of greenstone, and the schists of Table I, which outcrop near the syenites.

DULUTH GABBRO

After reaching such a conviction as to the origin of granites, Winchell found similar evidences that the basic rocks of the Keweenawan were derived from the metamorphism and complete refusion of the greenstones (in which he included sediments which may be Lower Huronian).[26] The evidences are grouped as (1) structural field relations, (2) petrologic and petrographic, and (3) general considerations.

The field evidence was said to be the passage of true gabbro to "granulitic gabbro," and this to greenstone by insensible gradations. The "granulitic gabbro" has a hornfels texture, identical with those contact rocks described in recent texts as granoblastic. The gabbro in a few places has a texture that approaches this somewhat, but the writer can only say that having visited hundreds of contacts of the gabbro in search of basal segregations, he has found scarcely one outcrop which left any doubt as to whether the rock was gabbro, or an exomorphic contact rock. Grant in 1900 had a similar experience.[27]

The petrographic evidences cited were very general. The greenstones, contact rocks, and gabbro all show variations. Although the contact rocks have minerals characteristically rounded or in a mosaic, they do in places show poikilitic textures and may possibly develop laths of plagioclase in augite, like the diabasic phases of gabbro. This close approach to gabbro textures is very rare, however, and the poikilitic arrangement of other minerals is a sign of contact action rather than of fusion. The mineralogic resemblance between recrystallized basalt flows and a gabbro can hardly be called evidence that the gabbro is derived from refusion of the flows.

Winchell's third, or "general considerations," referred to the

[26] N. H. Winchell, *Minn. Geol. and Nat. Hist. Surv. Final Rept.*, Vol. V, pp. 70–73 and 979–87. A. N. Winchell agreed to this evidence in *Amer. Geologist*, Vol. XXVI (1900), p. 157.

[27] U. S. Grant, "Contact Metamorphism of a Basic Igneous Rock," *Geol. Soc. Amer. Bull.*, Vol. XI (1900), p. 509.

petrographic variations in each formation, especially to some quartz-gabbro, which has since been shown almost certainly to be due to assimilation.[28]

Winchell clearly saw that if the gabbro was formed of fused greenstone, it should mark the place of an earlier greenstone belt.[29] He attempted to show that it does, saying that between Gunflint Lake and Kawishiwi River the gabbro is in contact with the green stone continuously; and from there southwest it lies along the granite that has intruded greenstone.

Several changes should be made in this outline of the relation of gabbro to greenstone. The gabbro evidently has a floor.[30] This makes it necessary to find a source of heat within the mass itself, for neither the roof nor the floor was fused. It also makes it necessary to find a source of magmatic substance (if not in profound depths) in the rocks that overlay the present floor. The roof of the gabbro for 130 miles is Keweenawan trap and felsite flows, apparently derived from the same source as the gabbro itself. The rocks now constituting the floor are the best guides we have as to the nature of material that could have been fused in place to make the magma. The floor mentioned by Winchell (see preceding paragraph) includes in "greenstone" such a variety of formations that a complete revision is needed. The rocks available are as tabulated.

Miles	
For 90	Virginia, Rove, and Carlton slate
15	Knife Lake slate (and Ogishke conglomerate)
15	Biwabik and Gunflint iron formation
20	Giants Range granite
5	Ely greenstone

Analyses of these formations show the improbability of deriving gabbro from such material as is available. The belief that the gabbro was so formed was probably based largely on the development all along the gabbro of contact rocks that resemble gabbro in color

[28] Frank F. Grout, "The Probable Extent of Abyssal Assimilation," *ibid.*, Vol. XLI (1930), pp. 678–87.

[29] *Op. cit.*, pp. 72–73.

[30] Frank F. Grout, "The Lopolith, an Igneous Form Exemplified by the Duluth Gabbro," *Amer. Jour. Sci.*, Vol. XLVI, p. 516.

and luster. The mineral composition of these rocks is not easily seen in the field and where they are derived from greenstone they have actually the minerals of gabbro. The much more abundant sediments, however, are not capable of yielding gabbro by fusion. Compare Analysis 8, computed as the average of the floor, with Analyses

TABLE V

ANALYSES OF DULUTH GABBRO AND ADJACENT FORMATIONS

	Adjacent Formations							Average	Typical Gabbro		
	1	2	3	4	5	6	7	8	9	10	11
SiO_2	58.04	64.77	58.45	61.39	63.92	71.45	52.46	62.13	48.20	46.45	47.90
Al_2O_3	18.66	14.45	18.99	16.97	None	12.99	16.77	14.90	19.53	21.30	19.92
Fe_2O_3	1.51	1.84	1.42	.39	31.13	2.44	3.92	4.72	Trace	.81	4.92
FeO	5.98	4.54	6.72	5.32	3.13	1.78	5.44	4.87	10.60	9.57	9.78
MgO	3.24	2.34	3.05	3.84	.49	.84	5.20	2.54	9.28	7.90	4.55
CaO	1.02	2.33	1.25	3.21	None	.65	6.23	1.59	8.51	9.83	8.56
Na_2O	2.12	1.37	2.96	2.78	Trace	3.03	2.58	2.13	2.52	2.14	2.75
K_2O	3.28	5.03	2.28	1.25	Trace	4.79	1.39	3.03	.32	.34	.56
H_2O+	3.28	1.92	3.82	2.44	1.12	1.24	3.28	2.49	.65	1.02	.76
H_2O-	.38	.07	.....	.06	.48	.10		.17	.08	.14	.....
CO_2	.14	.41	.47	.88	None	N.d.	2.14	.37	.02	.....	.....
TiO_2	.71	.60	.12	.62	Trace	.13	.45	.39	.65	1.19	.57
ZrO_2	.13	.....	.....	.07	.....	.10	.....	.05	.....	.....	.....
P_2O_5	.12	.20	.....	.19	.05	.04	.16	.11	.19	.02	.....
SO_3	.70	.60	.....	.....	.....	.....	.....	.27	.....	.....	.....
S	.17	.....	.....	.15	.....	.....	.....	.05	.03	.....	.....
MnO	.08	.11	.....	.12	.10	.04	.....	.07	.14	Trace	.....
BaO	.44	.....	.....	.06	.....	.....	.....	.10	.....	.....	.....
Cr_2O_3	.....	.....	.....	.01	.....	.02	.....	Trace	Trace	.....	.....
Carbon, etc.	.56	.....	.....	.....	.....	.....	.....	.12	.....	.....	.....
Others	.....	.....	.63	.....	.....	.....	.....	.13	Trace	.04	Trace
	100.56	100.58	100.16	99.75	100.42	99.64	100.02	100.23	100.72	100.75	100.27

1. Virginia slate; drill core from Sec. 1, T. 57 N., R. 17 W., depth 88–1723 feet. GEORGE WARD, analyst.
2. Rove slate, near Gunflint Lake. T. M. CHATARD, analyst. *U.S. Geol. Surv. Bull. 695*, p. 618, Analysis E.
3. Carlton slate; composite of several samples near Carlton, Minn. Analyses supplied by A. W. JOHNSTON.
4. Knife Lake, graywacke slate. Type locality. GROUT and HARTWELL, analysts. See also Table II.
5. Iron formation, Sec. 28, T. 58 N., R. 17 W. GEORGE STEIGER, analyst. *U.S. Geol. Surv. Mono. 43*, pp. 139–40.
6. Giants Range granite, biotite phase (see Table IV).
7. Ely greenstone; average (see Table IV).
8. Average of Analyses 1–7 inclusive in volumes estimated above. Compare this with the following.
9. Duluth gabbro with olivine; main phase at type locality. F. F. GROUT, analyst. See *Jour. Geol.*, Vol. XXVI (1918), p. 647.
10. Duluth gabbro, representative, near east end of Mesabi Range. H. N. STOKES, analyst.
11. Duluth gabbro with olivine, east of Birch Lake. A. N. WINCHELL, analyst.

9, 10, and 11. Probably the most conspicuous discrepancies are found in the excess of potash and silica and in the shortage of lime and magnesia in the floor rocks. Note also that, in the gabbro, iron oxide is mostly in the ferrous state and alumina is relatively high even though the slates are apparently aluminous rocks. The magma was no doubt intruded as gabbro magma from a deep-seated source.

SUMMARY

Minnesota has been supposed to show petrographic gradations into granite from graywacke, quartzite, conglomerate, arkose, and mica schist; also to quartz porphyry from sediment. The gradation may be admitted, but is now considered a result of injection and a little assimilation, not of metamorphism of sediments to the point of fusion. The compositions recently determined show that even aqueo-igneous fusion could not give the igneous rocks found.

The great Duluth gabbro has been similarly attributed to fusion of older rocks in place, but the available rocks have been analyzed and are not of the right composition.

It is noteworthy that the two groups of rocks attributed to fusion show very different discrepancies in composition. To make a granite from the schists involves additions of silica and alkalies. To make a gabbro from the rocks available involves not only a reduction of silica and potash, but large additions of lime and magnesia.

Assimilation of walls and xenoliths is evident in both granite and gabbro, but no proof of important magmatic changes due to this cause has been found.

Modern theories as to the deep sources of magmas and their general course of differentiation seem to be fully supported by the new studies of compositions of Minnesota rocks.

SUGGESTIONS OF FUSION IN THE EARTH'S 10-MILE CRUST

Although the conclusions from recent studies in Minnesota and those of most writers in recent years are opposed to the idea of a "hydrothermal fusion" producing igneous rocks almost in place, the idea of fusion in the crust of the earth is revived so frequently and so persistently that it may be well, if possible, to rescue from the controversy whatever germ of truth there may be in it.

The idea is an old one, having been presented by Hutton in 1785, if not before his time. Winchell gave an abstract of the discussion for one hundred years,[31] and it is clear that the leaders in geology up to 1880 commonly interpreted granite and gneiss as wholly or

[31] N. H. Winchell, *Minn. Geol. and Nat. Hist. Surv. Final Rept.*, Vol. V (1900), pp. 22–27.

partly fused sediment.[32] N. H. Winchell himself wrote a number of papers,[33] even discussing as a theme[34] the proposition that all acid eruptives are fused siliceous sediments. Since 1890 most writers have avoided the statement of fusion,[35] but some apparently consider fusion an extreme of the more readily acceptable processes of assimilation and contact metamorphism.

The discussion is complicated by the notion (maintained through about the same years) that granite could result from a schist or gneiss by progressive metamorphism. Many writers postulated the formation of granite with this theory in mind, without making any specific reference to fusion or to magma. In the Lake Superior region, Irving, Wadsworth, and Lawson in the decade 1880–90 showed that the granites and gneisses were clearly of igneous origin and younger than the schists above them. The truly igneous nature of granite has been increasingly accepted until the notion of its being made of sedimentary rocks through deep burial was said by Barrell,[36] in 1918, to be as obsolete as the still older doctrine of the Neptunists that granite was a crystalline precipitate on the floor of a primitive ocean. The metamorphic theory of granite is a "retrogression" in geologic theory. Granites are formed by magmas, though possibly some magmas may replace earlier rocks.

The question still remains as to whether magmas ever form in the earth's 10-mile crust or only at greater depths. A goodly company of geologists have presented some evidences and arguments in recent years to indicate that rocks may be melted at such shallow depths that later erosion may expose them to the light of day.

Daly doubted the origin of basaltic magma by any fusion of ordi-

[32] J. D. Dana, *Amer. Jour. Sci.*, Vol. VI (1873), p. 164; Bayley, *op. cit.;* Lawson, *op. cit.*

[33] *Amer. Geologist*, July, 1897, p. 41; November, 1898, p. 299; *Lake Sup. Min. Inst.*, August 22–24, 1911, p. 69.

[34] "Some Thoughts on Eruptive Rocks of Minnesota," *Amer. Assoc. Adv. Sci. Proc. for 1888* (1889), p. 219.

[35] In 1895 C. R. Keyes (*U.S. Geol. Surv. 15th Ann. Rept.*, p. 731) referred to the theory that granites like those in Maryland are fused sediments, but he was then opposed to that theory.

[36] J. Barrell, "The Growth of Knowledge of Earth Structure," *Amer. Jour. Sci.*, Vol. XLVI (1918), pp. 147–48.

nary sedimentary or gneissic material,[37] but has advocated the formation of a carbonate-rich magma from a dolomite.[38] Emerson saw evidences of fusion in Massachusetts,[39] but withheld conclusions for lack of analyses. Holmes reports evidences of fusion on a small scale where silicic rocks are heated by basic magmas, and quotes Thomas and Hawkes for similar facts. He then concludes that large bodies also may be fused in the crust by circulating basic magma.[40] Lane believes that some small amounts of hydrous pegmatitic magma may be generated locally in the crust.[41] Harker described a field occurrence[42] that is almost conclusive as to fusion. Sederholm is one of the most ardent advocates of fusion in the crust and has introduced the term "granitization," which seems to include some examples of the formation of granite magma from a conglomerate.[43] Quirke has described a whole series of rocks in which it is doubtful whether the igneous portions were added or were being formed and drained out.[44] Holmquist reports areas in Sweden where there seems to have been "regional pegmatization without connection with any granite,"[45] but adds that he finds no proof of regional refusion. Schwarz finds some South African rocks which "support the aqueo-

[37] *Igneous Rocks and Their Origin* (1914), p. 174.

[38] "Carbonate Dikes of the Premier Diamond Mine," *Jour. Geol.*, Vol. XXXIII (1925), pp. 683–84.

[39] B. K. Emerson, "Difference in Batholithic Granites According to Depth of Erosion," *Geol. Soc. Amer.*, Vol. X (1899), pp. 499–500.

[40] A. Holmes, "Association of Acid and Basic Rocks in Central Complexes," *Geol. Mag.*, Vol. LXVIII (1931), p. 248.

[41] A. C. Lane, "Origin of Granites . . . . by Selective Solution," *Bull. Geol. Soc. Amer.*, Vol. XXIV (1913), p. 704.

[42] A. Harker, "The Tertiary Igneous Rocks of Skye," *Mem. Geol. Surv.* (1904), p. 245.

[43] J. J. Sederholm, "On Regional Granitization (or Anatexis)," *12th Internat. Geol. Congress*, pp. 319–23.

[44] T. T. Quirke, "Killarney Gneisses and Migmatites," *Bull. Geol. Soc. Amer.*, Vol. XXXVIII (1927), pp. 753–70; "Structures and Batholiths of the French River Area," *Jour. Geol.*, Vol. XXVII (1929), pp. 698–99.

T. T. Quirke and W. H. Collins, "The Disappearance of the Huronian," *Can. Geol. Surv. Mem. 160* (1930), pp. 50–55, 102.

[45] P. J. Holmquist, "Swedish Archean Structures and Their Meaning," *Bull. Geol. Inst. Upsala*, Vol. XV (1916), pp. 125–48.

igneous solution theory of rock magmas,"[46] but his descriptions indicate only contact action and assimilation which are usually considered quite distinct from magma formation. Finally, Eskola in a recent paper on the origin of granite magmas supports the idea of differential anatexis during orogeny,[47] but the partial fusion he describes probably occurs at depths near 50 kilometers.

On the other side of the argument Leith and Mead note[48] that no case of fusion has been proved, and that chemical evidence in the few cases tested has proved unfavorable.

In the Huronian areas described by Quirke, analyses show that the rocks that have been "granitized" (whether by fusion or injection) show a considerable increase in potash content over the sedimentary originals. The writer made a somewhat hasty search for data on the chemical results of deep-seated contact action on sediments and found nine pairs of analyses from a wide range of districts. In summary of the results, it seems that for a few feet from the magma the sediment may have its potash content raised about five-tenths of 1 per cent—say from 2.50 to 3.00 per cent. The changes in the rocks described by Quirke are of a quite different order of magnitude, large masses having been changed from 1.39 to 4.50 per cent K_2O. It seems very doubtful if fusion of sediments even with contact emanations from some deeper magma could change the potash content so far as that. More likely there were large additions of the magma itself. The problem is an especially difficult one in this Huronian area because of the variety of gneisses present, some resulting from piezo-crystallization, others from *lit-par-lit* injection, and still others from a metamorphism clearly later than the latest igneous pegmatites.

SATISFACTORY ILLUSTRATIONS OF FUSION IN THE CRUST

The temperatures of magmas are believed to range from 1,200° C. in the basalts of crater lakes (where part of the heat may be from burning gases) down to 500° C. for the late pegmatites. The gabbro

[46] Contribution to the Aqueo-igneous Solution Theory of Rock Magmas, *Geol. Soc. of S. Africa, Trans.*, Vol. XV (1912), pp. 115–23.

[47] P. Eskola, "On the Origin of Granite Magmas," *Min. und Pet. Mitth.*, Vol. XLII (1932), pp. 455–81.

[48] C. K. Leith and W. J. Mead, *Metamorphic Geology* (1915), pp. 268–70.

magmas are thus more likely than others to produce a high temperature in their walls and inclusions and it is here that we should look for fused rocks in the crust. Bayley naturally supposed that the heat of the gabbro on Pigeon Point was sufficient to heat thoroughly and fuse its roof. That roof, however, was quartzite and perhaps less fusible than some other sediments.

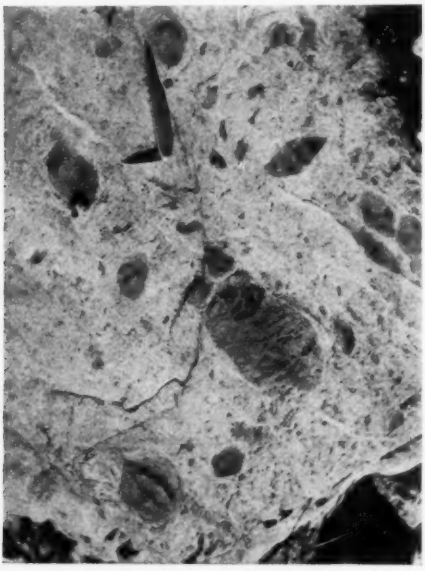

FIG. 2.—Hornfels of Animikie slate near the Duluth gabbro. In this specimen and many places along the contact, fragments of hornfels lie in a matrix that is also hornfels. The matrix was probably at least partly melted and could flow.

The writer has tested several of the formations along the north side of the Duluth gabbro in Minnesota by dropping small fragments into a muffle furnace at 1,150°–1,200° C. Both the slates and greenstones soften and swell viscously at such temperatures. If the gabbro had such temperatures when intruded, it is to be expected that certain fragments might be melted. The fusion of shales in drilling operations is also significant.[49] We may well conclude that

[49] N. L. Bowen and M. Aurousseau, "Fusion of Sedimentary Rocks in Drill Holes," *Bull. Geol. Soc. Amer.*, Vol. XXXIV (1923), pp. 431–39.

the temperature at some places in the crust is high enough to melt some rocks.

In the field and petrographic study of the inclusions in the gabbro several features support this somewhat theoretical conclusion. The hornfels texture in the contact rocks is very characteristic and not to be confused with ordinary igneous textures. The texture is found in some places, however, in such a relation as to suggest that the material was "mushy" if not wholly liquid; (1) hornfels dikes cross hornfels inclusions; (2) hornfels fragments may be cemented by hornfels into a breccia (Fig. 2); and (3) hornfels may grade into a

FIG. 3.—Magnetite dikes cutting hornfels inclusions in the Duluth gabbro, near Iron Lake, Cook County. The inclusion is so closely associated with altered iron formation that it may well be suggested that the magnetite dikes result from heating the iron oxides of that sedimentary formation.

diabase, though the gradation has not been found in such exposures as to prove which rock changed to the other. A final field argument for fusion may be based on magnetite dikes of small size crossing hornfels inclusions in the gabbro (Fig. 3). They occur near some fragments of iron formation, and it is well known that magnetite might melt at a lower temperature than the siliceous rocks that make up most of the formation. To be sure, the gabbro has formed some segregated magnetite bodies, but they are not very much like the dikes here mentioned.[50]

Turning to the rocks which have been heated only by the general earth heat, or by siliceous magmas at lower temperatures than gabbro magmas, we find only a few evidences of fusion in a small way

[50] T. M. Broderick, "Relation of the Titaniferous Magmetites of Northeastern Minnesota to the Duluth Gabbro," *Econ. Geol.*, Vol. XII (1917), pp. 690–96.

in the crust. The writer has seen (at Saganaga Lake in Minnesota) clear inclusions of greenstone in a granite. Along the margins of one inclusion a few feet across the greenstone appears to grade into a small lamprophyr dike and the dike intrudes and includes fragments of granite. No other explanation than fusion of the greenstone can

FIG. 4.—Lamprophyr dike in a conglomerate that forms the contact rock of a stock at Snowbank Lake, Minn. Note that although the dike cuts the conglomerate, the matrix of the conglomerate seems to flow into cracks in the fractured dike.

be considered as probable without much more information than is now available.

At Snowbank Lake, Minnesota, near a syenite stock, a lamprophyr cuts the conglomerate wall of the stock and appears to be clearly disrupted by the matrix of the conglomerate. See Figure 4, which shows quite clearly how the matrix of the conglomerate seems to have injected the dark dike which injected the conglomerate. This is perhaps analogous to some of the conglomerates found by Sederholm injecting other rocks.

Finally, the writer has seen sedimentary formations folded in the

deep zone, and in the outcrops where a sharp fold grades into a fault, and friction was probably most intense, the rock had locally a little pegmatite with all the petrographic character of an igneous rock. This apparently corresponds to the local pegmatites that Lane discovered. In rocks such as are shown in Figure 5 the outcrop rarely

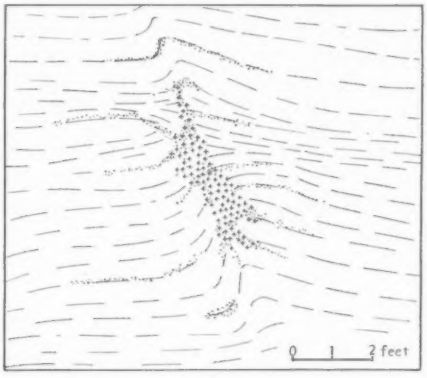

FIG. 5.—Sketch of a fold in sediments metamorphosed at great depths, now exposed on the south shore of Pickerel River, Key Harbor area, north of Lake Huron, Ont. Where the fold is so sharp that the beds broke, there is a little pegmatite with definitely igneous character.

shows three dimensions, and it may yet be asked whether the pegmatite might not have been injected from below as an apophysis from some large body at no great distance.

CONCLUSIONS

Thus all the examples of fusion in the crust that come to mind prove to be small bodies; and, furthermore, all are close to some large body of igneous rock that came to its place from a distance, It may be suspected, therefore, that the large magma contributed to the formation of the small magmas in the crust not only heat but mineralizers. These examples, though fairly conclusive evidences of fusion, do not serve to suggest any great importance in the process of magma generation in the earth's crust—that portion of the lithosphere so near the surface that later erosion might bring it to the light of day.

If there are any areas where extensive quartzites have been impregnated and replaced by magmatic emanations, or later by the magma itself, so that a great mass of rock assumes all the characters of a granite and would therefore be classified as an igneous rock, the areas should be given immediate and careful study.

Recognizing that the volume of a magma may be somewhat increased by solution of the crust, it still remains probable that there is very little fusion, as distinct from solution. Most magmas appear to reach the crust from below.

REVIEWS

Coral Reefs and Atolls. A course of lectures delivered at the Lowell Institute at Boston, 1930. By J. STANLEY GARDINER, F.R.S., F.R.G.S., London: Macmillan & Co., 1931. Pp. 181; pls. 15, figs. 33, map at end.

The appearance of a volume on *Coral Reefs and Atolls*, by the professor of zoölogy and comparative anatomy at Cambridge University, is an event of no little importance to students of reef formations. For the greater part of a lifetime Dr. Gardiner has been a student of reef formations and he has taken a leading part in a number of important expeditions for their study; and few men now living have seen and studied so many coral formations. He is, moreover, the only survivor of that group of eminent zoölogists which launched an early and high-powered offensive against the Darwin-Dana theory of origin of pacific atolls. One is, therefore, not surprised by Gardiner's somewhat extravagant praise of the fairness of view and the accuracy of description of coral reefs by the late Alexander Agassiz.

The book is well organized with chapters on "Fringing, Barrier and Atoll Reefs," "Island Formations," "The Natural History of Corals," "Other Plant and Animal Reef Builders," "The Other Organisms of Reefs," "The Distribution of Coral Reefs," "Atolls and Their Lagoons," and "The Foundations of Atolls." A most valuable Appendix of 15 pages is given over to terse descriptions of the principal reef groups which have been studied, and there is a very full Index. The illustrations are mainly new and include well-chosen photographs by Colonel Sewell and the author.

The earlier chapters of the books, which deal with the characteristics of the different types of reef formation and of their natural history, are naturally the most successful. The colorful beauty of coral formations lends itself to fascinating word pictures, and the literary skill of Professor Gardiner has here been used to good advantage.

For students of geology the great interest of the volume is naturally to be found in the later chapters, where the origin of atolls is gone into. Here Dr. Gardiner by certain expressions betrays his unfamiliarity with geological topics, and one is not surprised that facts which are favorable

to the subsidence theory of origin have been left out of consideration. It is unfortunate that, in common with some other students who have set out to explain the tectonics of atoll formation, Dr. Gardiner should have lumped together all coral structures of whatever sort. In addition to the atolls of the mid-Pacific area, which Darwin and Dana studied and which are each an individual truncated cone rising from profound depths of the sea, there are the arcuate mountain ranges with their associated reef formations which rise in the sea off the margins of the Asiatic continent and its extension southeastward; and again there are the great platforms crowned with relatively low reef pedestals (Bahamas, Yucatan Bank, Sunda and Sahul platforms, etc.). Lastly, there is the great area of recently sunk and later elevated sea floor of the western Indian Ocean, which has been Dr. Gardiner's special province.

To assume that the same tectonic movements can explain these widely different reef formations is to start out with an absurdity, yet Gardiner treats the tectonics of "the Indo-Pacific equatorial belt," which cuts athwart examples of all of these types, in order to prove an "average elevation" of 8–10 feet—although in the Fiji Islands there have been reef elevations of a thousand feet and more, and in the outer Pacific groups subsidences of many times that amount.

It was Marshall who first drew attention to the significant sharp line which outlines truly oceanic islands of the Pacific, and the subsidence theory of Darwin and Dana was applied to this oceanic area. Along its margin of growing mountain arcs, it is characteristic of the manner of evolution of a growing anticline that coastal terraces, veneered with fringing reef coral, show elevation upon the outer or convex side of the arc at the same time that barrier reefs and shallow atolls testify to subsidence upon the inner or concave side. Since the zoölogists who oppose the theory of subsidence have accounted for atolls by elevation of a submerged mountain crest through accretion of layers of deep-sea limestone, they have generally chosen for their examples not the atolls but mountain arcs for proof that coral formations undergo elevation—thus Guppy in the Solomons, Gardiner and Agassiz in the Fijis, and Semper in the Pelews. The "Tertiary limestones," so often reported by Agassiz from observation through binoculars while cruising past Pacific islands, have again and again been proved to be fringing reefs only which were veneered over volcanic tufaceous deposits. "Tertiary limestone" basements of atolls are today quite unknown, though this argument is used by Gardiner in his book.

In 1910 Wood Jones presented a very effective picture of the way in which boring plants and animals remove coral and other reef material

and transform it into the soft "coral mud" which covers much of the lagoon floor, as well as the seaward slopes of the atolls down to the depth of a mile or more. Even more convincingly is this subject presented by Gardiner out of a wider and more intimate experience. It is hardly to be doubted that these organisms are largely responsible for the constant shaping of the lagoon, though the depth of the floor is doubtless fixed by the wave base of the greatest storm waves in the lagoon. Obviously, no proof is here supplied that the lagoon of an atoll does not fill up during subsidence. Yet Gardiner dismisses the subsidence theory of atoll formation, saying: "It is superfluous to detail all the arguments opposed to the original Darwin theory, ... , since one fact kills it, viz., that there is no such general filling in of lagoons by coral growths and by sediment as suggested."

On the contrary, the green alga *Halimeda*, as Gardiner well emphasizes, grows luxuriantly within the lagoon as well as outside it down to a depth of 40–60 fathoms; and the 200-foot boring made within the lagoon of Funafuti passed, not through coral mud, but through *Halimeda* to its bottom. Even more significant, this organism was encountered to the extreme depth of 1,140 feet in the main boring through the outer Funafuti reef, though dredgings outside showed it only to a depth of some 60 fathoms. That contributions of reef material are made to the lagoon through the attacks of storm waves upon the reefs themselves is proved by the collection of great "nigger heads" and lesser débris seldom absent from the back slope of the windward reefs.

When atolls are elevated, the extension of the reef's islets toward each other to reduce the inlets into the lagoon, combined with the copious tropical daily rain, prevents the generally low tide of coral seas from replenishing the saline constituents within the lagoon. The water thus becomes brackish and eventually fresh, with the result that rookeries of sea birds become established with guano deposits and mangrove swamps. Under these conditions, and these only, the water of the lagoon becomes acid and solvent for the reef formation. Gardiner has applied these data as though they were generally applicable to the atoll areas of the mid-Pacific. His examples have all been drawn from the area of the western Indian Ocean, where such consequences of uplift are best displayed and available from the work of Fryer, Gardiner, and others. The elevation is always clearly indicated through the shallowness of the lagoon, and frequently, also, by the solution channels in the rock, as well as by the cliffs outside. The shallowing lagoon and the ultimate residual "sinks" of elevated phosphate islands have been treated by Gardiner in reverse order

in an effort to prove that solution had produced the lagoon basin—his discussion deals, therefore, with a devolution instead of an evolution.

Opponents of the subsidence theory of atoll formation are required to assume a centrifugal growth of the atoll reefs in order to explain the reef material of the main Funafuti boring through a supposed migration of the reef outward over its own talus. The point is one much disputed, and no wholly satisfactory data seem to be available. Gardiner, however, cites a comparison of measurements a half-century apart made on the rising atoll of Diego Garcia in the Indian Ocean, where the area of the northern part of the lagoon appears to have been enlarged during the interval by about one-eighth of its area, and the linear dimensions proportionally. This change, however, seems to be fully explained by the increased emergence of the frustum of the cone made by the reef pedestal and the simultaneous movement of the surf line down the outer slope, rather than by any considerable outward growth of talus. Such an effect is in fact inevitable in the case of a rising island, to which in the case of Diego Garcia the shallow lagoon of only 17 fathoms and the nearly continuous rim of islets bear testimony.

Despite its one-sided treatment of the origin of the atolls, the book is a most valuable contribution to the knowledge of the subject, and is particularly valuable in all that relates to the natural history of the reef organisms.

W. H. HOBBS

The Baltic Region—a Study in Physical and Human Geography. By E. G. WOODS. New York: E. P. Dutton & Co., 1932. Pp. 434; maps and sketches 143.

This excellent book by Mrs. Woods illustrates the increasing use made of geological data in an up-to-date geographical text. In each region the author outlines the geological structure of the older rocks forming the floor of the area, and this is followed by a more detailed treatment of the deposits of the Quaternary period. Thus the geologist will find a much more complete study of the effects of the advancing and retreating ice in the Baltic area than is usually considered advisable in a geography. For instance, the section on Fennoskandia consists of eight chapters of which two deal with pre-Quaternary geology, two with the Quaternary period, one with dislocation lines, and two with orography. The human aspect clearly occupies only a subordinate position. To the reviewer this is rather a welcome change from the large number of "human" geographies which omit the details of the basal structure of the region.

The author divides the Baltic region into three units. The ancient

Fennoskandian platform in the west and northeast, the Paleozoics, etc., of the east, and the Quaternary and Tertiary deposits of the south. Many geological maps and sections illustrate such features as the thrust of Kebnakaisse, the "windows" west of Sundsvoll, the fault blocks of Scania, and the distribution of "rhomb-porphyr." Few mining areas are more interesting than that around Gellivara in Lapland, where 270 million tons of rich iron ore occur. This has been exploited by the hydro-electric power at Porjus, where a fall of 245 feet gives 82,500 horse power. The widespread interests of a mining corporation are illustrated by the Falu Copper Company which dates from the thirteenth century. In addition to famous copper mines, it owns the largest steel works in Sweden, one of the largest paper mills in Europe, and the largest sawmills in the world.

In Finland there are good descriptions of the Salpausselka moraine, which extends for some 200 miles across the south, and of the shattered peneplain whose blocks emerge boldly from the clay covering near the coast, but are almost buried in the hinterland. An interesting account is given of the moors of Denmark and north Germany, with typical sections of the peat developed therein. The four chief end-moraines of North Germany are mapped and their relations to modern rivers and to the dunes are shown in several illustrations. Fluvio-glacial valleys formerly occupied almost the whole region between the middle Elbe and middle Vistula.

Between the Gulf of Finland and the Niemen River there is a more regular sequence of strata than is found elsewhere round the Baltic. Here a geosyncline seems to have been gradually filled during much of the time since the Cambrian with beds dipping to the south. Dolomites are common in the older beds, but there is no coal except some lignite in Kurland. Oil shale is found along the north coast of Estonia in Cambrian deposits. The whole area is, however, almost buried under deep morainic deposits.

The author has drawn freely on the large Swedish, Finnish, and German literature on the subject, and gives a Bibliography of many pages at the end of the book. A special feature of this unusual book is the large number of line drawings by K. C. Thomson, which in many cases emphasize the points of interest so much more satisfactorily than photographs would do.

GRIFFITH TAYLOR

Paleoklimatologie. By FRITZ KERNER-MARILAUN. Berlin: Gebrüder Borntrager, 1930. Pp. 510; figs. 21. M. 39 unbound.

This book will interest meteorologists as well as geologists, for it pursues meteorological methods as far as possible in determining past cli-

mates. Paleobotany also plays a major rôle in revealing past climates, as plants serve as thermometers and barometers from the Paleozoic to present. The probable distribution of dominant high- and low-pressure areas of the past is deduced by analogy from present conditions on the principle that similar conditions lead to similar results. These deductions are checked as far as possible by faunal, floral, and sedimentary evidence.

Large fresh-water bodies of the past are compared with our present Great Lakes in their influence on climate, and similar temperature-regulating effects are noted from the floras. Mountain ranges of the past and present are compared as agents of climatic control, and the effects of epicontinental seas are evaluated as far as possible.

The Pleistocene climates have been treated in many works, but the discussion has seldom had a background beyond the Pliocene; whereas, in this book, it has a retrospect of all of geologic time, and the treatment is unusually thorough.

J. T. McC.

Handbuch der Geophysik. By B. GUTENBERG. Berlin: Gebrüder Borntrager, 1930. Pp. 569; figs. 207. M. 70. (M. 52 by subscription.)

This is the third volume of a proposed ten-volume series on physical geology, the first two and the fourth and sixth being in press. Judging by this volume, the whole will be a monumental work on physiography; in fact, it takes in much which we normally place in structural geology, petrography, and petrology. About the only phases of geology not included are paleontology and economic geology. The discussion goes farther afield than is usual in our test and reference books; for instance, Wegener's theory of continental drift occupies more than 20 pages, and a discussion of the wandering of the poles a like amount.

There are chapters by Gutenberg on "Kraft in der Erdkruste," by von Wolff on "Plutonismus und Vulkanismus," by Born on "Erdkrustenbewegungen," by Gutenberg on "Geotektonische Hypothesen," and by Hess on "Mechanische Wirkung von Eis auf die Erdkruste." All of the volume is well written and excellently illustrated. The chapters on igneous action, vulcanism, crustal movements, and geotectonics will be of special interest to structural geologists. Physiographers will find this a valuable reference.

J. T. McC.